Bulletin
of
The Geological Society of America

VOLUME 58 November 1947 NUMBER 11

CONTENTS

PAGES

Flysch and molasse. By A. J. Eardley and Max G. White 979–990

Diagenesis of the Clinton hematite ores of New York. By H. L. Alling 991–1018

Archean geology of the Turku district in southwestern Finland. By Anna Hietanen 1019–1084

Subscription $10 per year.

Publication Office: Mt. Royal and Guilford Aves., Baltimore 2, Md.

Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117th Street, New York 27, N. Y.

NOTICE.—Four weeks notice of change of address is necessary. In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter in the Post-Office at Baltimore, Md., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103 Act of October 3, 1917, authorized on July 8, 1918.

PAPERS IN PRESS FOR FORTHCOMING ISSUES

GEOLOGY OF THE BORAH PEAK QUADRANGLE, IDAHO. By C. P. Ross

This paper is accompanied by a colored geologic map of the quadrangle. Separate copies of the map, unfolded and in tubes, may be secured. Price 25 cents apiece. Orders should be accompanied by remittance.

CORRELATION OF MISSISSIPPIAN FORMATIONS IN NORTH AMERICA. By the Mississippian Subcommittee, J. M. Weller, *Chairman*

PALEOGEOGRAPHIES OF SOUTH AMERICA. By L. G. Weeks

CENTRAL KANE COUNTY, UTAH: A GEOGRAPHIC AND GEOLOGICAL RECONNAISSANCE. By H. E. Gregory

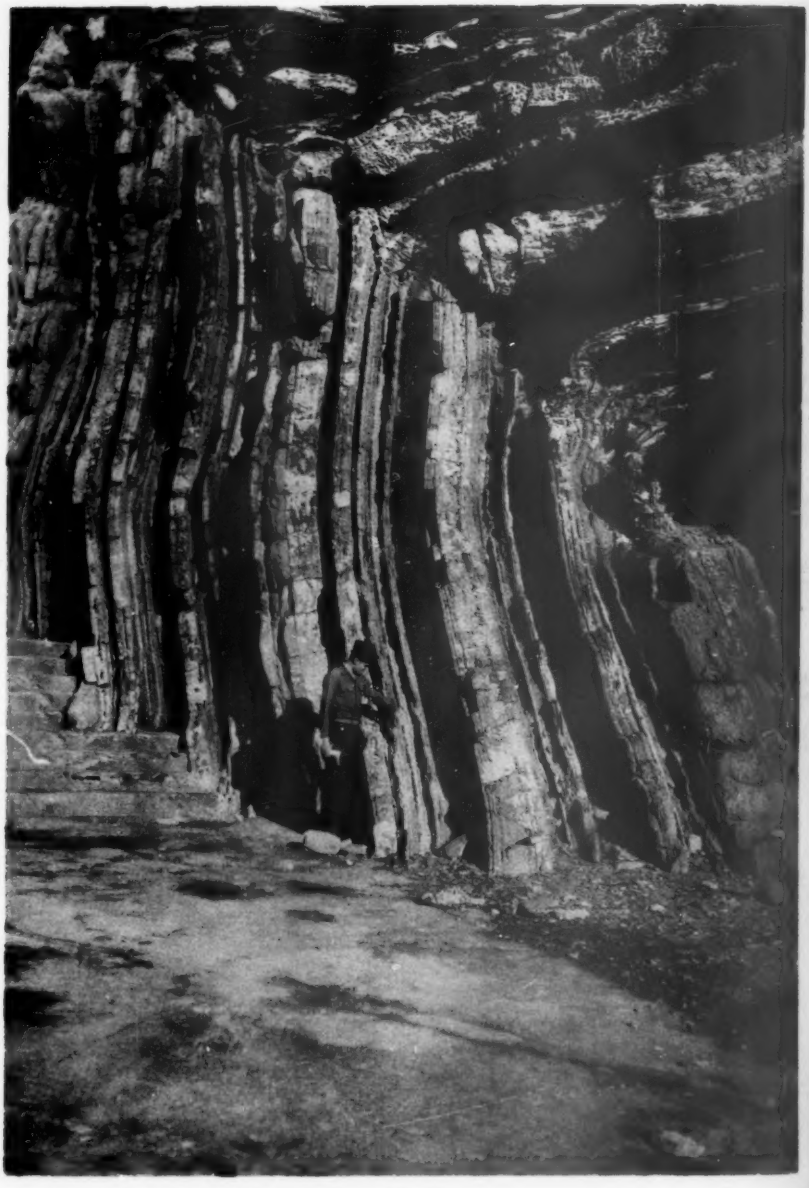
MINERALOGY AND PETROLOGY OF THE MAGNESITE DEPOSITS AND ASSOCIATED ROCKS OF CURRANT CREEK MAGNESITE AREA, WHITE PINE AND NYE COUNTIES, NEVADA. By G. T. Faust and Eugene Callaghan

GEOLOGICAL CORRELATION AND PALEOECOLOGY. By R. S. Allan

GEOLOGY OF THE TECOPA AREA, SOUTHEASTERN CALIFORNIA. By John Mason

This paper is accompanied by a colored geologic map of the area. Separate copies of the map, unfolded and in tubes, may be secured. Price 25 cents apiece. Orders should be accompanied by remittance.

CRITIQUE OF THE TIME STRATIGRAPHIC CONCEPT. By H. E. Wheeler



FLYSCH OF THE WESTERN PYRENEES

ALTERNATING MARLY LIMESTONES AND SHALES OF THE PIEDMONT OF THE WESTERN PYRENEES.
View of sea cliffs at east side of harbor of St. Jean de Luz.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 58, PP. 979-990, 1 PL., 1 FIG. NOVEMBER 1947

FLYSCH AND MOLASSE

BY A. J. EARDLEY AND MAX G. WHITE

CONTENTS

	Page
Abstract	979
Introduction	980
Flysch and molasse of the Alps	981
Original use of flysch	981
Type localities	981
Lithology	981
Thickness	982
Stratigraphy	982
Conditions of deposition	983
Swiss molasse	983
Flysch and molasse of the Pyrenees	984
Original use	984
Type localities and characteristics	984
Relation of Pyreneen flysch to Alpine	984
Difficulties in delimitation	984
Conditions of deposition	985
Molasse of the Pyrenees	985
Summary of European usage	985
Flysch and molasse of the Marathons	986
Flysch and molasse of the Taconics and Acadians	986
Difficulty with orogenic concept	987
Conclusions	988
References cited	988

ILLUSTRATIONS

Figure	Page
1. Flysch and molasse of the Alps	982
Plate	**Facing page**
1. Flysch of the western Pyrenees	979

ABSTRACT

By definition European geologists consider a sequence of limestones, sandstones, and shales, the beds of which are thin, regular, and alternating, and which are deposited in a geosyncline or foredeep shortly before a major orogeny, as the *flysch*. The waste products that accumulate as a deposit flanking mountains and built in part of the deformed flysch make up the *molasse*. In field practice the groups of sediments called flysch and molasse, or facies of them, are formations in the American sense. However, the Europeans would not recognize all groups of beds deposited in a geosyncline just before an orogeny as flysch; the beds must possess the proper lithologic and bedding characteristics.

Europeans have tacitly tied the variable of lithology and stratification to the variable of orogeny, with attendant difficulties. Originally only the first was denoted, but later the second was emphasized in definition if not in field use. Americans have considered the terms chiefly in their orogenic sense and thereby have called certain sequences flysch that do not fully meet the requirements of European usage.

The writers are of the opinion that little is gained by the use of the words; by their application no new fact is told or discovery made. They simply elaborate a conclusion by way of analogy. On the other hand, a possibility of confusion is introduced by the use of the terms, and in America it is best to avoid them.

INTRODUCTION

The terms flysch and molasse[1] are commonplace in European literature, and in the last 15 years they have been used by several Americans. After seeing certain European flysch and molasse sequences and investigating the literature on them, the writers are not satisfied that the meaning is clear and believe that in America the terms will not enhance our understanding of, or ability to interpret, an orogenic history.

In a very broad sense flysch indicates the sediments deposited in a sinking geosyncline shortly before a major orogeny. Van der Gracht (1931, p. 9), following European authors, defines flysch as:

> "a sequence of sediments deposited during the later stages of the geosyncline, directly previous to the major paroxysm, when initial diastrophism has already developed interior ridges exposed to erosion. In the . . . troughs of this early structure . . . a marine sequence of poorly fossiliferous clayey muds, with more or less sandy beds intercalated in the shales, was laid down to a great thickness."

He defines molasse as,

> "a generally much more clastic, very thick sequence, succeeding the Flysch; partly marine, partly deltaic, often containing enormous masses of conglomerate. It is the detritus worn down from the elevated ranges during and immediately succeeding the major diastrophism."

Primarily the terms have an orogenic significance; secondarily they imply a certain sedimentary character. The immediate difficulty lies in the fact that a sequence of beds may be called flysch on the basis of its lithology and stratification, and then the inference is made that it represents sediments that were deposited just before a major orogeny in a geosyncline. A great thickness and the location of the geosynclinal trough is thereby assumed, and perhaps on inadequate grounds. To certain European geologists flysch has solely a lithologic significance. To others it signifies an association—an alternating series—of certain lithologies. To others it is a mappable unit and therefore, in the American sense, a formation; the orogenic significance is largely neglected. To express the difference in another way, should we make sure that a geosyncline is indicated by the thickness, distribution, and composition of the strata, and then, if the lithology and stratification are correct, apply the name flysch; or is a certain association of strata flysch regardless of its orogenic relations?

---

[1] Most writers in English spell molasse with one l, but in French and German it is generally spelled with two ll's. Some Americans have capitalized the terms, but most have not. It is spelled with one l and not capitalized in the Oxford Dictionary.

A second difficulty is that the characteristic European flysch sequences are not well duplicated in the geosynclines in the United States where the term flysch has been used. Still a third difficulty, perhaps, comes from geologic settings in which a series of compressional episodes have followed in succession, and two or three successive molasses are laid down, without each being preceded by flysch deposition.

After a brief discussion of the flysch and molasse of prominent localities in Europe and of sequences in America to which the terms have been applied, the value and propriety of their further use in America will be summarized.

FLYSCH AND MOLASSE OF THE ALPS

ORIGINAL USE OF FLYSCH

The word flysch comes from the Swiss-German dialect, and means "gliding rocks" (Gignoux, 1943, p. 499). It was originally used by Swiss quarrymen as a trade name for particularly well-bedded, easily quarried sedimentary strata used for building stone. Geologists early applied the name flysch to the rocks in which the quarries were located (Lapadu-Harguess, University of Paris, oral communication).

TYPE LOCALITIES

The type flysch is in the western High Alps, in France and Switzerland. The strata there have been studied and described in detail. The flysch of the Pyrenees is almost as well known as that of the Alps and was originally called flysch because its lithology and stratification reminded an early geologist of the Alpine flysch. It is not the same age, is probably much thinner, and is not exactly similar lithologically to the Alpine flysch.

LITHOLOGY

The Alpine flysch consists typically of alternating layers of thin-bedded, hard, fine-grained, limey sandstone, limestone, and dark-gray clay, shale, or slate (Gignoux, 1943, p. 499). The beds range in thickness from a fraction of an inch to several inches. If one lithology becomes dominant an adjective is added such as *flysch noir*, for an abundance of dark shales, *flysch grêseux*, for a sandy flysch, and *flysch calcaire* for a flysch in which beds of limestone 30 feet or more thick occur (Fig. 1). A thick part of the Alpine flysch is made up of argillaceous limestone with intercolations of shale or clay. The tracks or imprints of *Helminthoidea labyrinthica* (thought to be either an annelid or marine mollusk) are abundant on the bedding surfaces, and the section is called the *flysch à Helminthoides*. *Wildflysch* is a conglomeratic or breccia facies of the flysch in which giant boulders 10 to 70 feet or more in length are imbedded in the shales and limestones. The *wildflysch* is also referred to as a *cordilleira facies* and is thought to be a submarine landslip or the collapse debris of submarine cliffs at the time of deposition of the flysch, or to be associated with the early growth of the great nappes. The Alpine *wildflysch* is probbably similar to parts of our Chaney shale of the Ouachita Mountains and to the great boulder bed in the Haymond formation of the Marathons.

THICKNESS

No precise data were found anywhere in the literature on the thickness of the flysch. A few writers refer to it as hundreds to thousands of feet thick. There can be no doubt that it is very thick and of geosynclinal proportions. The great Niesen nappe of Switzerland is composed almost entirely of it, and it is intimately involved in the structure elsewhere of the High Alps.

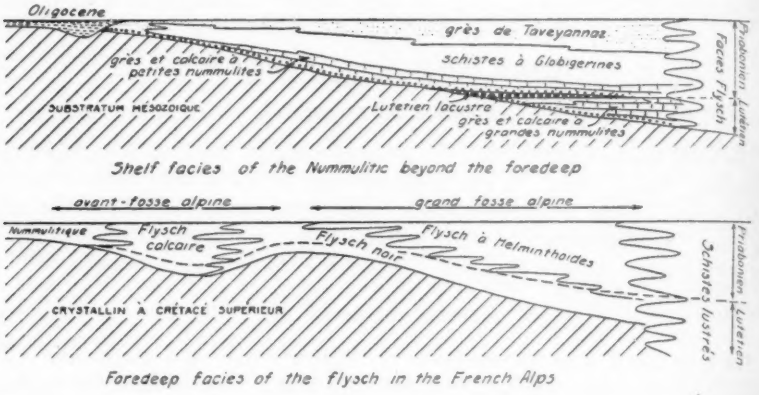


FIGURE 1.—*Flysch and molasse of the Alps*

Idealized sections restoring the Alpine flysch of the foredeep and the equivalent Nummulitic of the shelf beyond the foredeep to the conditions of deposition before deformation. After Gignoux (1943, p. 498 and 502). The upper section should be imagined reduced in vertical dimension so that the "facies flysch" on the right will match the "Flysch calcaire" on the left of the lower section. The differentiated *Nummulitique* of the upper section is the undifferentiated *Nummulitique* of the lower.

STRATIGRAPHY

The flysch of the High Alps contains few fossils, but some have been found. They are sufficient to correlate it approximately with thinner formations to the north, northwest, and west in a peripheral belt. The strata of the peripheral belt are very fossiliferous and are of two stages, the Priabonian (upper Eocene) and the Lutetian (middle Eocene). These deposits are transgressive toward the west, *i.e.*, away from the thick flysch, with the Lutetian resting unconformably on the Cretaceous and the Priabonian overlapping the Lutetian (Fig. 1). The Lutetian and Priabonian beds represent the Eocene portion of the Nummulitic and may be traced from the Mediterranean coast along the Alpine front into Switzerland and beyond. They are thought to represent deposition in shelf areas in front of the foredeep in which the flysch was deposited.

The Alpine flysch itself is so varied lithologically and so complex structurally that its units have not all been arranged in stratigraphic succession, nor have they all been dated. Although most of it is Eocene, some think the flysch contains Cretaceous beds: others think the lower Eocene is absent as it is in the shelf area; still others believe that some members of the flysch may be Jurassic or Cretaceous (Gignoux, 1943, p. 500).

CONDITIONS OF DEPOSITION

The flysch of the Alps was deposited in the foredeep of the rising hinterland, here largely of deformed Cretaceous strata. The sandstone facies at the top of the flysch sequence marks the final filling up of the foredeep at the beginning of the compressional deformation. The orogeny of the hinterland had now spread into the foredeep where the sediments were compressed and folded into great *nappes de charriage*.

It is interesting to note that only the Eocene of foredeep facies is called the flysch in the Alps. There are great thicknesses of marls, limestones, and shales of Jurassic and Cretaceous age, also referred to as trough deposits, but these are not called flysch. They can be distinguished from strata called flysch by color and lithologic differences such as the *terres-noires*, a black shale and limestone, or by metamorphism, such as the *schist lustre*, for the most part a shiny phyllite. A number of lithologic units in the geosynclinal Jurassic are spoken of as facies, and the *facies dauphinois* is specifically called a foredeep deposit by Gignoux (1943, p. 354). It is "thousands of meters" thick and composed mostly of dark shales and limestones. From Gignoux's excellent treatise, the term flysch is to him simply a convenient facies name, like various other facies names in the Alps. An American geologist would probably think of them as formations with their various members or units.

SWISS MOLASSE

The word molasse was first used as an adjective by Brongniart (1827) to describe a loosely cemented building stone in Switzerland. It was adopted a few years later by D'Omalius d'Halloy (1833) as a geologic term and still later again used by Coquand (1857).

The typical molasse is described by Cayeux (1929, p. 166) as a calcareous arkose, predominantly light gray, coarse-grained, friable, and very porous. It is a sandstone with much feldspar, some mica, chlorite, and glauconite, and cemented by marl or calcium carbonate. It rarely contains fossils.

In the Swiss Basin the lowest molasse is Oligocene. It is a deltaic deposit with intercalated lenses of conglomerate and thickens and coarsens toward the mountains. The name *Nagelfluh* is applied to the conglomerates.

All the Miocene of Switzerland is molasse. The section has three parts: a lower fresh-water unit, a marine unit, and an upper fresh-water unit. The lower fresh-water unit of the section is a typical gray molasse, *viz.*, a sandstone with high feldspar content and calcite cement. The marine unit is a glauconitic arkosic sandstone with cementing material generally composed of calcite with some clay. It is generally light gray but in places is yellowish or brownish. The upper fresh-water unit is a micaceous, chloritic, greenish arkose with calcareous cement. No statement of the precise thickness of the Swiss molasse was found, but it is implied that it is several hundred feet thick.

The arkosic nature of the Swiss molasse is taken to indicate that the clastics were laid down after the appearance of the large nappes in the Alps and after their sedimentary cover had been removed. The grantitic, crystalline core of the nappes alone could serve as source material for the amount of feldspar in the molasse.

An example of the use of molasse and flysch, seemingly with formational sense, is quoted from Collet (1927, p. 88):

"....Then we see the Wildhorn Nappe thrust over the Molasse (Tertiary of the Swiss Plateau). But between the thrust plane of the Wildhorn Nappe and the Molasse we find the *subalpine Flysch*, a very complicated formation,...."

FLYSCH AND MOLASSE OF THE PYRENEES

ORIGINAL USE

Stuart-Menteath (1894) was the first to recognize the similarity of a series of sedimentary rocks in the western Pyrenees to the flysch of the Alps and named it accordingly. Almost all workers in the western Pyrenees since have continued the use of the term flysch, and others have extended it to the eastern Pyrenees.

TYPE LOCALITIES AND CHARACTERISTICS

The specific locality of the flysch of Stuart-Menteath is in the Pays Basque of France and Spain. Intensive studies of the flysch have since been made in certain areas of the Pays Basque of Spain by Lamarre (1936) and of the Hendaye region of France by de Lapparent (1918). The flysch of the Hendaye region is an alternating sequence of regularly and thin-bedded shale, marl, and limestone. The individual beds range from 3 to 18 inches in thickness, and chert lenses, beds, and nodules are a conspicuous part of the limestone. Plate 1, is an example of the flysch at St. Jean de Luz. No sandstone is present, and no silt or sand content is obvious in the shale, marl, or limestone.

The total thickness of the typical flysch in this region is not given in the literature, probably because it cannot be determined anywhere with reasonable assurance. The writers have measured flysch in excess of 500 feet and believe that it may exceed 1000 feet in thickness but are persuaded that it is not of so-called geosynclinal thickness (Eardley, 1946, p. 1529).

RELATION OF PYRENEEN FLYSCH TO ALPINE

The lithology of characteristic sequences in the two places is approximately the same. Marls and marly limestones are the dominant lithologies. The Alps have a sandy facies, the Pyrenees a cherty. Thin and regular bedding is common to both.

The Pyreneen molasse is the same age as the Alpine flysch, but the Pyreneen flysch is appreciably older than the Alpine flysch. The Alpine flysch is thousands of meters thick, the Pyreneen flysch perhaps not much more than 1000 feet.

DIFFICULTIES IN DELIMITATION

The flysch of the Pyreneen region is probably all of Turonien and Lower Senonien age (Gignoux, 1943, p. 415). This typical flysch is overlain by about 1500 feet of massive marl of Upper Senonien, Danien, and Paleocene age, which separates it from the Nummulitic of Eocene and Lower Oligocene age which in part has been called molasse. A formation of nonflysch type between beds called flysch and molasse does not occur in the Alps. If an orogenic-sedimentation principle is implied by

the Alpine flysch-molasse sequence, then here the generalization has a conspicuous exception.

CONDITIONS OF DEPOSITION

The flysch of the western Pyrenees crops out in a piedmont zone today, and the conditions of its deposition are not so clear as in the Alps. A rising hinterland has been postulated in the site of the Pyrenees, but the mountains there are made up chiefly of Hercynian crystallines overlain by massive and thick Upper Cretaceous limestone. The massive, thick limestones are the same age as the flysch and could not have been a highland to the flysch trough. The position of the foredeep, in fact the existence of a foredeep, has not been demonstrated.

It seems clear that the term flysch as used in the western Pyrenees should mean solely a certain lithologic assemblage. Until the meaning of a regularly bedded alternating sequence of shales, marls, limestones, and sandstones is well established, it is not wise to infer much regarding orogeny.

MOLASSE OF THE PYRENEES

The molasse of the Aquitaine Basin is distributed along the eastern portion of the basin from the Gironde estuary to the Pyrenees foothills. It is principally of Oligocene age, but Eocene beds have been identified in it. It is a continental deposit equivalent in age to the upper part of the marine Nummulitic, and consists of sandy marl, siltstone, clay, and sandstone, in part arkosic, with calcareous cement. Toward the south a conglomerate facies of the molasse appears and becomes thicker and coarser toward the Pyrenees. This has been called *poudingue*. Fresh-water limestone intercalations appear in the region of the Massif Central. Nine hundred feet of molasse is known from a well in the vicinity of Agen (Mengaud, 1926), and Eardley (1946) has measured 2500 feet of Nummulitic near Biarritz.

The molasse of the Aquitaine Basin was derived largely from the thrust sheets of the Pyrenees. Its argillaceous and sandy nature has led to the belief that either a greater admixture of Pyreneen sedimentary rocks went into the formation of the molasse there than in the Swiss molasse or that considerable reworking of the material took place in the basin while sediments from surrounding lowlands were added.

SUMMARY OF EUROPEAN USAGE

Contained in the European usage are two variables—lithology and stratification, and orogeny. Originally only the first was denoted, but as the structure of the Alps became better understood the sedimentary characteristics were associated with the closing stages of geosynclinal subsidence and the deformational paroxysms that followed. It has not been demonstrated, however, that sedimentary conditions necessary to produce a characteristic flysch exist only in the foredeeps and just prior to the folding and thrusting of the foredeep sediments. Where the flysch and molasse concept was first transferred—the western Pyrenees—neither the required sedimentary environment nor the orogenic chronology is demonstrable.

With this insecure footing it is interesting to note what Americans have done with the terms.

FLYSCH AND MOLASSE OF THE MARATHONS

The Pennsylvanian rocks of the Marathon region, west Texas,

"are predominantly clastic, made up of sandstone and shale, with some beds of limestone and conglomerate. They reach a thickness of at least 12,000 feet in the southeastern part of the area but are much thinner to the northwest. The lowest formation, the Tesnus, is a mass of shales and fine sandstones 6,500 feet or more thick; the Dimple is a limestone as much as 1,000 feet thick; the Haymond formation, 3,000 feet thick, is again sandstone and shale, with a remarkable boulder bed in the upper part. The upper 1,800 feet of the succession, the Gaptank formation, is an alternation of limestones and conglomerates, with sandy and shaly beds. Marine fossils are found sparingly in the Dimple limestone and parts of the Haymond formation but are most abundant at numerous horizons in the Gaptank formation. Fossil plants, mostly abraded and poorly preserved, have been found at several places in the Tesnus and Haymond sandstones." (King, 1937, p. 55)

The Haymond formation is noted not only for its boulder bed but for its regular, rhythmic layering of sandstone and carbonaceous shale. The layers are a fraction of an inch to several inches in thickness. There are a few thicker sandstone beds and near the base some thick shales. At a few places the sandstones and shales have thin intercalations of sandy limestone, some of which contain marine invertebrate fossils. The upper part of the formation contains thick layers of massive arkose, and in the syncline east of Haymond station there are several members of boulder-bearing mudstone as much as 150 feet thick. The boulders in the mudstone beds include erratic blocks of the older formations as much as 130 feet across. According to King (1937, p. 87):

"The character of the Tesnus and Haymond sedimentary rocks and their relation to the diastrophic history is comparable to that of the Flysch of Alpine chains in Europe.

"Both the Tesnus and the Haymond formations appear to have been laid down under shallow water, for the bedding surfaces of many of the sandstones are channeled and ripple-marked, and thin layers persist for long distances in the exposures, as if the sediments had been well sorted and distributed by waves and currents before they were deposited. Whether the water in which the sediments were laid down was marine is uncertain, for great thicknesses of strata are quite barren of fossils, which if present would give some clue to the nature of their environment.

"Beginning with the upper part of the Haymond formation there was a great change in sedimentation. Most of the older clastic rocks were derived from the erosion of crystalline rocks beyond the limits of the Llanoria geosyncline. From this time onward the greater part of the fragments consisted of older Paleozoic rocks of the Llanoria geosyncline itself.

"The conglomerates in the upper part of the Pennsylvanian section indicates nearby contemporaneous uplifts of considerable magnitude, although no unconformities have been found in the areas of sedimentation. At the top of the Pennsylvanian section is a great unconformity by which it is separated from the Permian rocks that overlie it on the northwest. This unconformity marks the culmination of the orogeny, but the record of the sedimentary beds beneath it suggests that a considerable part of the folding and faulting in the Marathon Basin had taken place earlier, while deposition was still going on in the region.

"The younger Pennsylvanian strata are comparable to the Molasse of the Alpine ranges."

FLYSCH AND MOLASSE OF THE TACONICS AND ACADIANS

Marine carbonate deposition gave way to terrigenous argillaceous and arenaceous accumulations in the late Ordovician in New York and Pennsylvania, and with the change a more rapid subsidence of the geosyncline set in. Over 2000 feet of gray shales, gray sandy shales, and red shales was deposited over a large area. The Taconic disturbance then occurred, and the early Silurian terrigenous sediments, in places over 6000 feet thick, overlapped those of late Ordovician.

"Whereas Ordovician sediment progressively coarsened, the Silurian is successively finer and less extensive; they form the 'flysche' and 'molasse' of the Taconic revolution" (Kay, 1942, p. 1620).

The shales of the Upper Ordovician are not much like the Alpine flysch either in lithology or stratification, and it is very doubtful if a European geologist would call them flysch, although they fit the orogenic relations.

Kay (1942, p. 1632) also likens the sediments that accumulated prior to and after the Acadian disturbance to flysch and molasse. He says:

> "The next great influx of terrigenous detritus was in the middle Devonian—; the later rocks generally coarsen upward and westward, the coarser and non-marine facies gradationally overlapping the finer and marine. The sediments form the flysch from rising highlands to the east, whose uplift culminated in the Shickshockian (Acadian) disturbance; they are succeeded by Mississippian molasse."

If only the orogenic relations are implied, Kay's usage is correct, but again by European practice there is little in lithology and stratification of the Upper Devonian to resemble Alpine flysch.

DIFFICULTY WITH OROGENIC CONCEPT

In parts of the American Cordillera a series of orogenic stages have followed in close or fairly close succession, so that the molasse of an earlier orogeny was deformed and in part furnished the sediment for a later molasse. In the Wasatch Mountains of Utah the Frontier formation would be considered geosynclinal flysch to the later molasse which is the Henefer formation (Eardley, 1944, p. 840). However, the Henefer was much disturbed by the Willard thrusting, after which the coarse sediments of the Wasatch group accumulated. Both Henefer and Wasatch are molasse, but the Wasatch was not preceded by the deposition of flysch.

Again in the Hoback Range and Hoback Basin of northwestern Wyoming the geosynclinal flysch of the first Laramide disturbance is not found in the immediate hinterland. The molasse is the Hoback formation, some 15,000 feet of shales, arkosic sandstones, and a few conglomerate layers. The molasse is lower Eocene, but the next oldest sediments are those of the Frontier of middle Upper Cretaceous age. A long time separates the two during which the flysch should have been accumulating. The Hoback was overthrust by Mesozoic and Paleozoic strata, and a new molasse was deposited, the Pass Peak conglomerate of middle Eocene age. The setting is actually more complicated than implied by the foregoing description because in nearby regions several formations of uppermost Cretaceous and Paleocene age exist, and the spatial as well as the time relations are complex.

The geologic history of western Nevada and California during Triassic, Jurassic, and Cretaceous time is one of great complexity. Triassic and Lower, Middle, and early Upper Jurassic sediments in geosynclinal thicknesses were deposited east of a highland that contained many volcanoes. Then the sediments were intensely folded and injected with great batholiths to become a highland. A new trough formed on the west where the former highland stood and in turn received a great thickness of sediments. Most of these came from a third land area that had risen still farther west. Volcanic contributions were very abundant in most of the formations, and it would be difficult to apply the terms flysch and molasse to these sequences. The European geologist who has mapped the flysch of the Alps or the Pyrenees as a formation would see little in common.

CONCLUSIONS

The terms flysch and molasse have been defined by Europeans to designate bodies of sediment with certain lithologies and related to orogeny, but in actual practice they are used more as formational names. Since the flysch and molasse of one orogenic belt may not be the same age as the flysch and molasse of another the use of a common formational name may lead to confusion. Also, since a flysch must accumulate in the foredeep and just before the main paroxysms, it should first be established that such is the geologic setting, but certain of the well-known sedimentary sequences have been called flysch solely from the lithologic and bedding characteristics. It is confusing also that one deposit of Nummulitic is the foreland shelf facies of the flysch and another near by is part of the molasse. Again, the flysch is separated from the molasse in one prominent European locality by a third formation, and this is confusing if it is implied that a certain orogenic succession of events is always the rule, and that this is always recorded by a certain sedimentary assemblage. The mistake lies in tying two variables together in a fixed relation.

The few geologists who have applied the terms to deposits in North America have followed the orogenic requirements of the definition and made certain that the sedimentary sequences were the required associates of an orogeny. In the Marathons, the lithologic character matches the European fairly well, but it is doubtful if Europeans would wish to apply their terms to rocks in the Acadians and Taconics because the lithologies there are somewhat different from those to which they are accustomed.

Also in the American Cordillera the orogenic and sedimentary history has been one of time and spatial complexity in such a manner as not to follow exactly the pattern of growth of the Alps. Particularly it is clear that no flysch appears to accompany a molasse in one place, that several molasse one after another develop as a result of a succession of orogenic disturbances in other places, and that volcanism and shifting highlands and troughs result in thick orogenic deposits that are hard to differeniate on the basis of flysch and molasse.

The writers are of the opinion that little is gained by the use of the words in many places; by the application of the terms no new fact is told or discovery made. They simply elaborate a conclusion by way of analogy. If there is a possibility of confusion by the use of the terms it is best to avoid them.

REFERENCES CITED

Boussac, Jean (1911) *Nummulitique de Biarritz*, Ann. Hébert, Ann. Stratigr. Paleont., Faculté Sciences Univ. Paris, Tome V.

Brongniart, A. (1827) *Classification et carectères minéralogique des roches homogénes et hétérogénes.*

Cayeux, L. (1929) *Les roches sédimentaires de France, roches siliceuses, Carte Géol. France, Mém.*

Collet, Léon W. (1927) *The structure of the Alps*, Edward Arnold and Co., London, p. 289.

Coquand, H. (1857) *Traité des roches.*

de Lapparent, Jaques (1918) *Terrains Crétacés de la région d'Hendaye*, Carte Géol. détaillee France, Mem., p. 155.

D'Omalius D'Halloy (1833) *Introduction à géologie.*

Eardley, A. J. (1944) *Geology of the north-central Wasatch Mountains, Utah*, Geol. Soc. Am., Bull., vol. 54, p. 819–894.

——— (1946) *Petroleum geology of Aquitaine Basin, France*, Am. Assoc. Petrol. Geol., Bull., vol. 30, p. 1517–1545.

Fournier, Eugene (1908) *Pyrénées Basques*, Carte Géol. France, Bull. Services, no. 121, Tome 18 (1907–1908).

Gignoux, M. (1943) *Géologie stratigraphique*, Mason & Cie., Paris.

Kay, G. M. (1942) *Development of the northern Allegheny synclinorium and adjoining regions*, Geol. Soc. Am., Bull., vol. 53, p. 1601–1658.

King, P. B. (1937) *Geology of the Marathon region, Texas*, U. S. Geol. Survey, Prof. Paper 187, p. 1–148.

Lamarre, Pierre (1936) *Recherches géologique dan les Pyrénées Basque d'Espagne, Soc. Géol. France*, Mem. 27, Nouvelle Série, Tome 12, p. 462.

Mengaud, L. (1926) *Sur quelques sondages profonds dans le Bassin de l'Aquitaine*, Soc. Hist. Nat. Toulouse, Bull., vol. 54.

Moret, L. (1934) *Sur la présence de* Nummulites incrassatus *dans le "Flysch Calcaire" des environsd' Orcières (nappe du Flysch de l'Embrunais)*, Soc. géol. France, C. re., 4 fév.

Stuart-Menteath, P. W. (1894) *Sur les lignes directrices des environs de l'Observatoire d'Abbadia (Basses-Pyrénées)*, B.S.G., C. re. sommaires.

Van der Gracht, W. A. J. M. van Waterschoot (1931) *The Permo-Carboniferous orogeny in the south central United States*, K. Acad. Wetensch. Amsterdam. Verh., Afd. Natuurk., deel 27.

University of Michigan, Ann Arbor, Michigan; U. S. Geol. Survey, Washington, D. C.
Manuscript Received by the Secretary of the Society, November 14, 1946

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 58, PP. 991-1018, 3 PLS., 9 FIGS. NOVEMBER 1947

DIAGENESIS OF THE CLINTON HEMATITE ORES OF NEW YORK

BY HAROLD L. ALLING

CONTENTS

	Page
Abstract	992
Introduction	992
Acknowledgments	992
Previous work	993
Clinton stratigraphy	994
Geological associations of the ores	995
Geological setting of the Furnaceville	996
Geological setting at Clinton, New York	998
Lenticular characteristics	999
Colors of the iron minerals	1000
Mineralogy	1001
Iron minerals	1001
Sulphides	1002
Chamosite	1002
Glauconite	1002
Francolite	1002
Carbonates	1002
Cementing minerals	1002
Leucoxene	1003
Minor minerals	1003
Dwarfed Fauna in the iron ores	1003
Theories for dwarfing	1003
Crinoid stems and hematite	1005
Chemical recasting	1005
General discussion	1005
Composition of oölitic hematite	1007
Types of ores	1007
Roundness of quartz nuclei	1007
Quantitative changes in forming the ore	1009
Source of the iron	1011
Diagenesis of Clinton iron ores	1012
Origin of Clinton ores	1012
Paragenesis	1013
Conclusions	1015
References cited	1015
References to chemical analyses	1017

ILLUSTRATIONS

Figure	Page
1. Stratigraphy of the New York Clinton. Based upon publications of Newland and Hartnagel, Chadwick, Sanford and especially Gillette	995
2. Chart of complementary microlithologies of the ore and associated rocks at Rochester, New York	998

3. Chart of complementary microlithologies of the ore and associated rocks at Clinton, New York 999
4. Margins of lenses of Furnaceville ore in the Rochester region 1000
5. Cross sections of bryozoan *Helopora fragilis*, showing size in nonhematitic limestone, in iron ores, and the fillings, replacements, and coatings. Somewhat conventionalized 1004
6. Normative composition of Clinton ores and changes in the formation of the ore from original limestone 1008
7. Size frequency-distribution curves of the two types of oölites 1009
8. Roundness and circularity of quartz grains in "eastern" type of ores 1010
9. Paragenesis of Clinton iron ores 1014

Plate
1. Fossil fragments in Clinton ores
2. Thin sections } At end of paper
3. Thin sections

ABSTRACT

Quantitative microscopic analyses from thin sections of the Clinton iron ores of New York and associated rocks reveal a succession of introductions and replacements which were responsible for formation of the ore. The iron was introduced by surface waters shortly after deposition and replaced a series of sedimentary minerals, chiefly carbonates, as part of the diagenesis of the rock. For this, the expression, diagenetic replacement, is here proposed.

Recently developed stratigraphic diagrams, employing the concept of "microlithologies", indicate that these ores were originally limestones which were replaced by iron compounds, now largely hematite, provided certain essential conditions prevailed and the clay content was less than 10 per cent by weight. The oölitic coatings consist of hematite and chamosite in alternating thin layers, onion fashion. The "fossil" ores are the result of the partial replacement of the carbonate of fossils and the subsequent oölitic coatings. These replaced fossils act as nuclei.

The Clinton ores are lenticular and possess little or no stratigraphic significance; they represent similar lithologic conditions.

INTRODUCTION

While the hematite iron ores of the New York Clinton are not now of economic importance, as a source of either iron or paint pigment, a discussion of their origin is of interest in itself, and particularly as a portion of the larger problem of the diagenesis of sedimentary rocks in general. This study furnishes much detail of the replacement of the original limestones by nonmagmatic iron-rich solutions and of the subsequent changes in the lithification of these rocks.

ACKNOWLEDGMENTS

Material for this study has been collected over many years. It began in 1923 with the field assistance of Merle K. Alling and has been supplemented from time to time by the co-operation of former students: Bernard H. Dollen, Virginia Hoyt Jones, William L. Grossman, and James S. Wishart. Through the courtesy of Albert O. Hayes of Rutgers University, Chris. A. Hartnagel and John G. Broughton of the New York State Museum, and James H. C. Martens of the West Virginia Survey, additional specimens have been made available. Thin sections were also loaned by Nelson C. Dale, of Clinton, New York.

The author used an unpublished manuscript of Miss Winnie McGlammery which deals with the dwarf fauna of the Furnaceville ore at Rochester. This manuscript is now in the library of the Department of Geology, The University of Rochester.

Correspondence with George H. Chadwick of Catskill, Charles K. Swartz of Baltimore, and Frank M. Swartz of State College, Pennsylvania, has been of great benefit.

The greatest assistance was rendered by my former student, Tracy Gillette, who became the leading authority on the New York Clinton. Our discussions and his published papers have been invaluable. His death in 1942 deprived us all of a remarkable geologist and a prince of a fellow. This paper is dedicated to him.

William F. Jenks, The University of Rochester, read the manuscript and made a number of valuable suggestions.

PREVIOUS WORK

James Hall (1843, p. 60) recognized that the hematite of the Furnaceville at Rochester replaced bryozoans and other calcareous fossils. He thought that the iron itself was derived from pyrite, which in turn replaced the shells after the strata were in their present position. Hall believed that the rock below "the drainage level" was largely limestone.

Foerste (1891) and Kimball (1891) advocated late replacement by hematite. On the other hand, Newberry (1880) maintained that the ore was an original deposit. C. W. Hayes (1894) concurred; he never found the ore beds in Tennessee passing into nonferruginous limestones. Diller (1898, p. 138) said that "the fossils [of the Furnaceville at Rochester] are chiefly broken crinoids and bryozoa. . . .Under the microscope . . . these fossils are, in most cases, found to be only partially made up of iron ore . . . By dissolving a fragment . . . in HCl . . . a considerable part of the $CaCO_3$ in the fossil fragments is seen to be replaced by silica."

Cayeux (1909, vol. I, p. 284–285) proposed that some of these ores were originally oölitic limestones. In a special study of the Clinton ores he advocated an elaborate series of replacements and alterations which A. O. Hayes (1915, p. 85) regarded as unwarranted. Cayeux proposed that it was first an oölitic limestone which was changed to oölitic siderite "then through chlorite to hematite. The siderite and hematite so derived gave rise to quartz, göethite, pyrite, etc."

Smyth (1911, p. 33–52) formerly held that these ores may have resulted from different processes at different places, but finally concluded "that with few exceptions, all the [types of the] Clinton ores must have" been formed by the same processes.

The replacement hypothesis as usually expressed implies that concentration of iron solutions in the original sediments was not marked but that ground waters circulating obtained iron from adjacent beds and replaced the limestone long after lithification.

Smyth believed the formation of the ores was primary probably by the deposition of limonite, which by dehydration and oxidation became hematite.

McCallie (1908, p. 185–194) proposed that the iron was originally deposited as glauconite instead of limonite. Today we should read "chamosite" for "glauconite".

Stose (1924, p. 409–411) reported that some of the ore from Wise County, Virginia,

is composed of iron carbonate. In the zone of circulating ground waters, this ore, he thought, could be oxidized to hydrated ferric oxide which through heat and pressure would become hematite. The carbonate of iron was secured by deep drilling, hence he did not wish to draw sweeping conclusions. He wrote: "There seems to be no reason to doubt that most Clinton hematite ores were deposited as ferric oxide, but the Virginia occurrence of carbonate ores seems to be one of the exceptional cases . . ."

A. O. Hayes (1929, p. 4) concluded that the Wabana "oölitic iron ore was an original sediment, and the hematite was formed before the consolidation of the rocks." But he also noted that the formation of the iron oxide itself is not yet definitely established. The iron oxide may have been precipitated from sea water, or it may be due to oxidation of precipitated chamosite. But if all the iron oxides came from chamosite the ores today should contain more silica than they do. Possibly some was derived in this manner. Hayes has decided that a solution to this problem "must await further study." He "thinks that some of the hematite is secondary after chamosite, and some was precipitated directly."

CLINTON STRATIGRAPHY

The present understanding of the stratigraphy of the New York Clinton is based upon the work of Newland and Hartnagel (1908), Chadwick (1918), Sanford (1935; 1936) and especially Gillette (1940; 1947). The interpretation of the diamond drilling by Newland and Hartnagel was the first modern report. They recognized three shales and two limestones below the Irondequoit limestone instead of two shales and one limestone as formerly supposed. Chadwick emphasized this discovery and named the newly recognized rock units. He also pointed out the unconformity between the (Lower) Sodus and the Williamson in the Rochester region. Sanford (1936, p. 798) further stressed the break in a suggestive diagram.

Gillette (1940; 1947) studied not only the stratigraphy and paleontology but also the accompanying lithologic changes and proposed a rational correlation of the western and eastern sections. Following Ulrich and Bassler (1922), he accepted the suggestion that some Middle Clinton occupied the unconformity. Chadwick (personal discussions, 1945–1947), however, feels that the establishment of the presence of "Middle Clinton" in New York State is not too certain and explains these relationships by facies changes. Gillette (1947) regarded the occurrence of the ostracod, *Mastigobolbina lata* in the Sauquoit beds as proof. The author accepts Gillette's conclusion.

That there are important facies changes in the Clinton of New York there can be no doubt. The following changes, from the base up, from west to east, are briefly noted:

(1) The Thorold sandstone changes into the base of the Oneida conglomerate.

(2) The Reynales limestone becomes more argillaceous and is known as the Bear Creek shale (of Gillette, not Chadwick), then with more siliceous material it constitutes a part of the Oneida.

(3) Both the Lower and Upper Sodus grade into a portion of the Oneida. The latter becomes younger to the east and in part becomes Middle Clinton in age.

(4) The Wolcott apparently pinches out or grades into the Sauquoit.
(5) The Williamson on becoming more siliceous is called the Willowvale.
(6) The Irondequoit acquires more quartz and takes on the facies of the Willowvale.

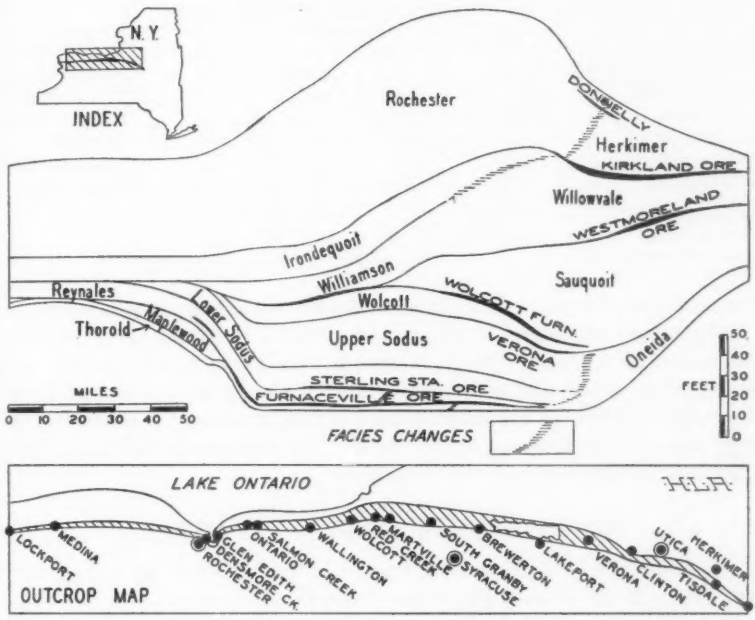


FIGURE 1.—*Stratigraphy of the New York Clinton*
Based upon publications of Newland and Hartnagel, Chadwick, Sanford and especially Gillette.

(7) The Rochester, with more siliceous minerals, grades into the Herkimer sandstone.

GEOLOGICAL ASSOCIATIONS OF THE ORES

The hematite ore beds extend some 150 miles east and west across the central portion of New York State. Usually only a few zones occur in any one locality. In the west the ore beds are associated with limestones and are locally termed "fossil" ores. In the east the rocks are slightly more siliceous although still highly calcareous, ranging from siliceous limestones to calcareous siltstones, some with shaly structure. These are usually the so-called "oölitic" ores. Newland (1908, p. 46) has stated that the two types are infrequently found together. It is obvious that even the "fossil" ores are essentially fossil fragments replaced by hematite and coated by oölitic layers of hematite-chamosite. In the "oölitic" ores the nuclei, instead of being fossil remains, are quartz grains of sand size. The reason for this difference is simple: fossil

nuclei occur in limestones, while the quartz nuclei are found in the more siliceous beds. As the latter occur in the eastern portion of the Clinton belt, the quartz-centered oölites are more abundant there. Both "fossil" and "oölitic" ores are really oölitic in structure.

TABLE 1.—*Stratigraphic classification of the ore zones*

Top of Rochester	*Donnelly Ore
Base of Rochester (Herkimer)	*Kirkland Ore
Base of Willowville	†Westmoreland Ore
Top of Wolcott	*Wolcott Furnace
Base of Wolcott	*Verona Ore
Top of Reynales	*Sterling Station
Base of Reynales	**Furnaceville

* Chadwick (1918).
† Gillette (1947).
** Hartnagel (1907).

The ore beds are long lenses and represent similar lithologic conditions. In Table 1 the various zones are listed. The names are those of Hartnagel (1907), Chadwick (1918) and Gillette (1947). They are used, not necessarily in their authors' sense, however, but as names for the ore beds without implying any specific stratigraphic rank. Gillette called the Furnaceville ore a formation because of its distinctive lithology and its great length along the strike. But other ores, such as the Sterling Station and the Verona were not so designated.

GEOLOGICAL SETTING OF THE FURNACEVILLE

At Rochester the hematitic rock is 1 foot to 14 inches thick and is essentially a "fossil ore." It lies between limy beds of the Reynales formation. The top and bottom contacts are relatively sharp and conformable. The Furnaceville contains sufficient calcite to effervesce with dilute hydrochloric acid. As the iron content is not extraordinarily high and the bed is relatively thin, it cannot compete commercially with richer and thicker layers. It was important in the past (Newland and Hartnagel, 1908, p. 54–55).

The method used in Figure 2, showing the composition of the rocks, has been described elsewhere (Alling, 1945, p. 737–756). The data which form the basis of the charts are quantitative microscopic measurements, Delésse-Rosiwal analyses, obtained by a Wentworth micrometric stage. These data have been recalculated into three (or four) complementary "microlithologies." This, in brief, was done as follows: the clay and the small quartz grains (less than 0.032 mm in nominal sectional diameter) were assigned to the argillaceous microlithology, the calcite and other carbonates, together with the indigenous chert, were placed in the calcareous class, while the coarse-grained quartz, feldspar, and the accessories (heavy minerals) were listed under the siliceous category.

There are, therefore, four columns to Figures 2 and 3, which, if laterally compressed would result in a single column with a total of 100 per cent (by weight).

TABLE 2.—*Tabulation of Clinton iron ores**

Locality	Donnelly ore	Kirkland ore	Westmoreland ore	Wolcott Furnace ore	Verona ore	Sterling Station ore	Furnaceville ore
Vanhornsville							
Tisdale		2′ 1″	several inches				
Willowville		4′ 2″	several feet				
Clinton		6′	2′ **2′ 1′			0	0
Verona	1′	6′		1′ 5″		1′	0
Lakeport				2′	1′ 10″	7″	4″
Brewerton				1′ 11″	trace	trace	1′ 4″
S. Granby				3″ **6′ 1″ 3″	several inches	0	8″
Martville						1″	3″
Red Creek				1′ 4″		4″	2′ 6″
Wolcott				1′		2″	2′ 9″
Wallington				2′ 10″ sh no hematite		1″	2′ 6″
Salmon & Second Creek				1′ 6″		trace	1′ 6″
Ontario							1′ 6″
Glen Edith							0
Densmore Ck							9″
Rochester							8″–14″
Medina							0
Lockport							0

*Data of Newland, Hartnagel and Gillette.
**Barren layer.

The section at Rochester is essentially composed of the argillaceous and the calcareous microlithologies, except for the hematite in the Furnaceville; the siliceous column could well be neglected. The siliceous column is required for the Clinton area where siliceous sediments and quartz-bearing flagstones occur.

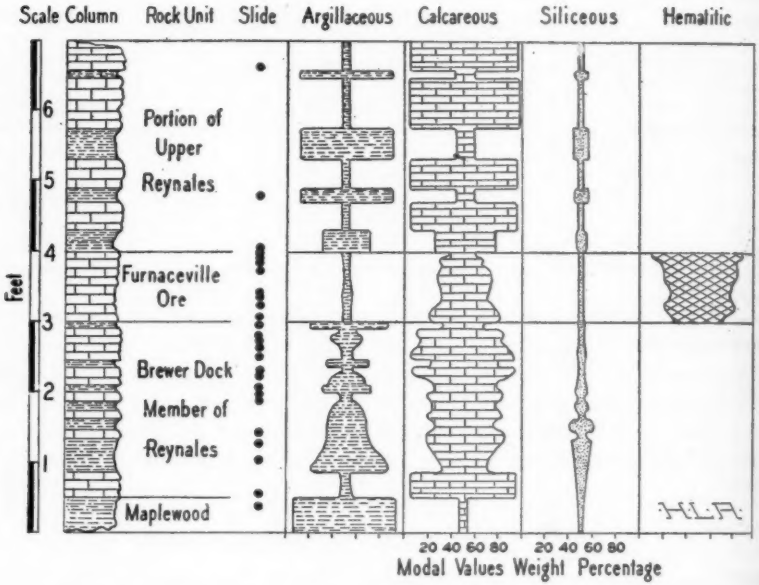


FIGURE 2.—*Chart of complementary microlithologies of the ore and associated rocks at Rochester, New York*

Showing the dominance of the calcareous microlithology.

The charts (Figs. 2, 3) show that those limestone beds which were replaced by hematite were low in the argillaceous microlithology, less than 10 per cent by weight. There are other limestones with a low clay content that appear to be equally suitable for replacement but are not hematite bearing. It is concluded that a low amount of argillaceous matter was one of the essential factors in the formation of these iron ores.

GEOLOGICAL SETTING AT CLINTON, NEW YORK

There are four beds of iron ores. The lowest is 1 foot thick, overlain by a 2-foot layer of siliceous rock. The upper oölitic bed is 2 feet thick. Eighteen feet higher is a thin layer of oölitic hematite; 3 feet higher is the 6-foot Kirkland ore or "red flux", so-called because the calcitic content makes it self-fluxing. The associated rocks have been described as "blue shale with thin sandstone layers" (Kemp, 1895, p. 104); they are shaly beds of calcareous and siliceous composition, with interbedded thin calcareous flagstones. Microscopic examination establishes that the ores were originally limestone layers.

LENTICULAR CHARACTERISTICS

From the early days of mining the ore beds have been known to be lenticular. Individual beds are not continuous over great distances, but the zones in which they

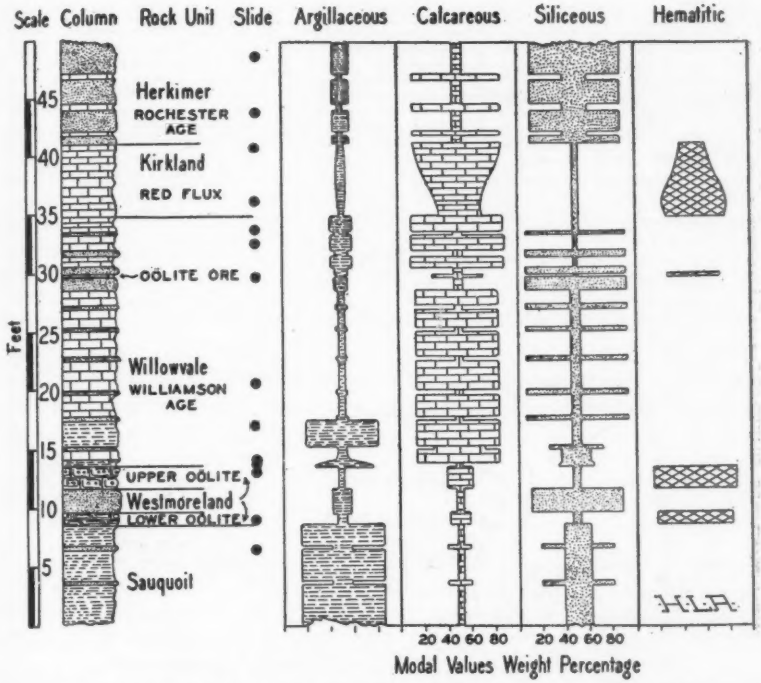


FIGURE 3.—*Chart of complementary microlithologies of ores and associated rocks at Clinton, New York*

Based upon specimens donated by Virginia Hoyt Jones, William L. Grossman, Chris A. Hartnagel, John G. Broughton, slides loaned by N. C. Dale, and those of the author.

This shows the dominance of limestones, though the rocks are more siliceous than at Rochester.

occur are much more extensive. One locality near Rochester shows this lack of continuity and was investigated.

The ore occurs in the Gorge of the Genesee and 3½ miles east of Rochester in Densmore Creek. But 2 miles farther east, on the east bank of Irondequoit Bay at Glen Edith, no hematite bearing rock occurs, even though the horizon is present. (*See* Table 2.) At Fruitland, 9 miles still farther to the east, the ore comes in again with this difference: the horizon is several feet lower, and is continuous to Ontario, Furnaceville and beyond. The latter town is the type locality for the Furnaceville. Actually the ore at Rochester is discontinuous and at a higher horizon than the Furnaceville ore. It could be argued that the ore at Rochester is not the Furnaceville and should be given another name but to do so is regarded as unwise and unnecessary.

The Furnaceville is defined in this paper as the ore that occurs at or near the base of the Reynales Limestone (Fig. 4).

The horizon at Glen Edith is characterized by cherty limestone with pyrite. At Fruitland the ore is in narrow ribbons with intervening nonfossiliferous limestone.

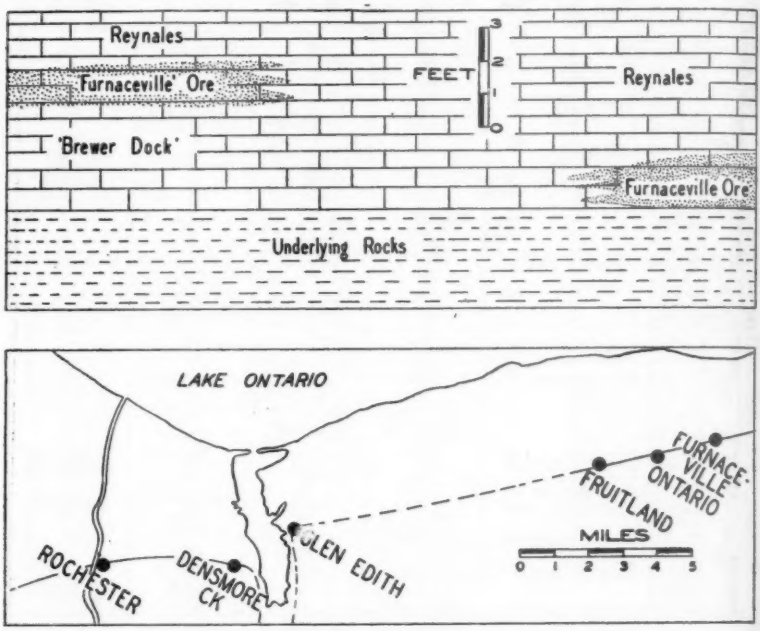


FIGURE 4.—*Margins of lenses of Furnaceville ore in the Rochester region*

The ore at Fruitland, Ontario, and Furnaceville is not continuous with that at Densmore Creek and in the Genesee Gorge at Rochester.

In this barren material are "inclusions" of hematite-replaced fossil fragments. The hematite is limited to water-worn fossil remains.

The exposures at Densmore and at Fruitland represent marginal phases of separate lenses of ore. Both seem to be characterized by more chert, especially at the top of the beds, than farther from the edge of the lenses. The chert appears to have been silica which replaced the carbonates, especially those comprising fossil fragments as a diagenetic process (White, 1947).

There is nothing to prove that the introduction of this silica was linked with the silica that was a part of the material of the chamosite usually associated with the iron compounds which are now hematite.

COLORS OF THE IRON MINERALS

Many attempts have been made to standardize description of rock colors. Some of the systems have been consulted, but were found inapplicable to the reds, browns,

and yellows of the iron-bearing minerals in these ores seen in thin section by reflected light. The lamps used were equipped with Corning daylight (pale blue) light filters. After some experimentation, the author adopted the colors of the colored pencils manufactured by the Joseph Dixon Crucible Company. The manufacturer's catalog numbers are listed so these can be duplicated by other investigators. The method,

TABLE 3.—*Color chart of minerals in iron ores*
(REFLECTED LIGHT)

Color	Dixon No.	Mineral Aggregate
Greenish buff	—	Chitin
Sepia	335	Francolite (Collophane)
Lake red	369	Early hematite
Tuscan red	426	Chamositic hematite
Terra cotta	397	Iron-stained clay
Raw umber	428	Ferric iron-stained francolite
Yellow	374	Ferrous iron-stained francolite
Burnt sienna	398	Ferric iron-stained chitin

of course, leaves much to be desired, yet it has much to commend it. A set of color charts was prepared and viewed with a camera lucida which permitted matching the colors directly to the minerals (Table 3).

MINERALOGY

IRON MINERALS

Studies of the Clinton ores in thin section by both transmitted and reflected light reveal that the iron minerals have many colors and structures. To some it is possible to assign definite mineral names; for others it is not justified. Hematite is the principal mineral. Turgite and goethite are also present in small amounts. But many grains appear to represent mixtures and iron-stained minerals other than hematite.

The chief structure, is, that of the oölites. These are alternating onion-skin layers of hematite and chamosite.

The nonoölitic hematite is brightly colored brick red, carmine, lake red, red, vermilion, and occasionally terra cotta. It is distinctly microcrystalline with sparkling platy surfaces. Its crystalline nature may well have been acquired after deposition.

The oölitic hematite involved with chamosite is duller and darker and is essentially cryptocrystalline. It contains some purple and is usually Tuscan red. Some of its darker hue is due to minute grains of martite identifiable under high magnification.

There are, in addition, a host of substances customarily called "limonite," the bulk of which is suspected to be clay and phosphatic materials, such as collophane (or its variant, francolite) iron-stained. Their colors range from sepia, burnt sienna, raw umber, orange, to lemon yellow.

Besides the lake-red microcrystalline hematite and the Tuscan-red cryptocrystalline oölitic chamosite-hematite, there is a dull-red or red-brown "powdery" form that is a submicrocrystalline mineral which appears to be high in hematite. It probably represents a mixture of several substances.

SULPHIDES

There are probably several ferrous sulphides, such as hydrotroilite and melrikovite. They are black, opaque, and difficult to distinguish from "carbonaceous matter." They seem to be the forerunners of much of the associated marcasite and pyrite, although the relations are not always clear. These black ferrous sulphides appear to antedate the pyrite in the majority of cases.

CHAMOSITE

This interesting mineral, a member of the chlorite group, is important in the formation of the oölitic structures in the ores. It seems very doubtful that they would have been formed without this gelatinous aluminosilicate. Both A. O. Hayes (1915, p. 58; 1931, p. 44–64) and Smyth (1892; 1894; 1911) stress its function. It almost always alternates with films and skins of the hematite, resulting in Tuscan red. It does occur separately, however, in narrow beds at Clinton, New York (Pl. 3, G, H, J).

Other chlorites are also present. Pennite occurs usually inside an oölite, interfingering with blood-red specularite.

GLAUCONITE

This mineral is commonly associated with hematite, usually surrounding it. Nothing was seen to relate its origin to biotite.

FRANCOLITE

The carbonate calcium phosphate is postulated rather than collophane. It is very common in the limestones rich in fossil fragments, hence it is probably akin to coprolites (Bradley, 1946).

CARBONATES

Several types of carbonates, usually calcitic, are recognized: (1) fossil fragments, some thoroughly recrystallized even though minute; (2) clastic carbonate aggregates of very small dirty grains which probably represent detrital fragments worn off from larger grains and chemically precipitated calcite; (3) recrystallized calcite, relatively low in magnesium; and (4) the calcite of crinoid fragments which contain a pale yellow-green tubular network of supposed chitin. (Pl. 1, D, G, H, J, K.) Alling (1946, p. 13) has referred to this as "netted calcite." Rarely is siderite recognized as a later development in the ores.

CEMENTING MINERALS

These Clinton rocks have been cemented by calcite and silica, now quartz. The latter has "enlarged" the clastic quartz grains. In the grain-size measurements this added material has been omitted, giving the size of the original grains in two dimensions.

LEUCOXENE

This white mineral is very abundant. Some grains have been sliced through the center showing the nature of the parent mineral. It is not always ilmenite, sphene, rutile, and brookite have been identified. In the ores, yellow to yellow-brown grains resulted from iron staining.

MINOR MINERALS

Clastic grains of zircon, tourmaline, corundum, and andalusite have been identified in the thin sections. They all show abrasion and are wellrounded, especially tourmaline. They probably represent several erosional cycles.

DWARFED FAUNA IN THE IRON ORES

THEORIES FOR DWARFING

The collection made by McGlammery contains many specimens of the bryozoan *Helopora fragilis*. These proved to be very instructive in thin section regarding the manner in which the mineral that eventually became hematite replaced the fossils. These byrozoans are round, double-pointed, needle-like fossils, 5 to 6 millimeters long and one half to three-quarters of a millimeter in diameter (Pl. 2, A–K). The zooecial apertures are oval in outline and arranged in twelve parallel rows so staggered that some seem to spiral around the stems. Cross sections show beneath some of the apertures additional cavities (Pl. 2 G). The zooecia appear to be composed of chitin, sometimes partially of calcite.

After the death of the organism, the byrozoans were apparently rolled by waves of the shallow seas, rounded and reduced in size. The apertures were filled by a variety of materials: clay minerals, francolite, clastic quartz, or calcite. It appears that the membranous chitin was stained by iron compounds and eventually replaced by lake-red hematite.

Many of these were photographed and copied by means of a camera lucida. Some of these, somewhat conventionalized are shown in Figure 5.

A thin section from near the base of the Reynales limestone from Lockport, New York represents the horizon of the Furnaceville ore, though it contains no hematite. There are many specimens of this bryozoan present. They are a little larger in diameter than those in the ore. They appear to have experienced less rounding by wave tumbling. It is possible to restore the original in dotted lines.

Paleontologists have long assumed that the high iron content of the Clinton sea was responsible for the dwarfing of these organisms, comparing them with the "Tully" fossils in the marcasite at the top of the Hamilton beds (Loomis, 1903). McGlammery (1931) stresses the biologic cause of the reduction. She gave measurements in support of this explanation.

Microscopic measurements were made on cross sections being careful not to include the oölitic coatings. As McGlammery's figures do not discount these additions, her values are not significant.

The actual reduction in size may have several causes: wave wear, chemical solution, as well as biologic dwarfing. It is probable that each species had its own degree of

reduction. Certainly the smaller forms are not exclusively the result of biologic dwarfing.

This 20 per cent reduction over the nonhematite limestone bryozoans is probably due to biologic causes.

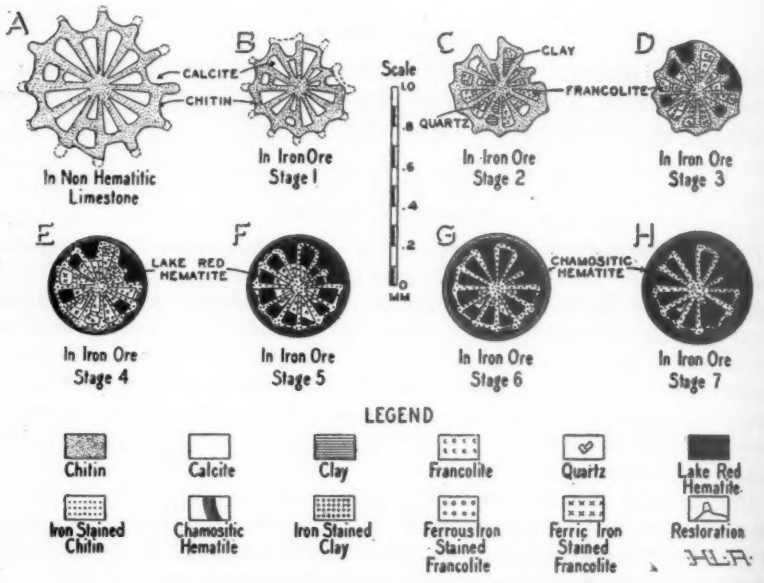


FIGURE 5.—*Cross sections of bryozoan,* Helopora fragilis

Showing size in nonhematitic limestone, in iron ores, and the fillings, replacements, and coatings in successive changes. Diagrams based upon camera-ucida drawings and photomicrographs, somewhat conventionalized.

Wesenberg and Medd (1901) reported that in the Danish Lake Furesco, the shells on the bottom occur in all stages from fresh and unchanged, through those coated with limonite to those completely replaced, even that some shells containing living organisms have been coated with iron compounds.

It has always been assumed that the cause of the dwarfed fossils was due to iron in the sea water. The author considers that it may have been due to an increase of salinity also. If this theory is correct, the iron entered these salty basins later than had been formerly supposed.

Clarke and Ruedemann (1903) have shown that the fossils of the Lockport dolomite, which occurs above the Clinton, were affected by an increase in the salinity of the gradually restricting sea which became the Salina. The fossils either developed large, coarse-walled shells or were small with thin walls. It seems clear that, if similar conditions obtained in the Furnaceville sea, the fragile shells could be easily broken. This suggestion interlocks the two observations: the fragmentation and the dwarfing of the fossils.

The observations on the fossil fragments show definitely that replacement was a dominant process, occurring before the development of the oölitic coatings, although there are a few exceptions to this order of events.

TABLE 4.—*Factors involved in measuring fossil reduction*

Cause	Reduction	Enlargement
Wave Wear	X	
Biologic dwarfing by salt	X	
Biologic dwarfing by iron	X	
Oölitic coatings		X

TABLE 5.—*Reduction of size, Helopora fragilis*

Diameter in ores. mm. Average	Diameter in nonhematite limestone	Same restored mm	Reduction Per cent	Number of specimens measured
.42	—	.56	25	47
—	.77	.81	5	16
		Difference	20	

The chitinous membrane was eventually replaced by the lake- or carmine-colored hematite. In some specimens, apparently where replacement has not gone quite so far, the color is deeper (Dixon No. 321½), resulting in a solid core of lake-red hematite Around this core has been deposited oölitic chamosite-bearing hematite of Tuscan red (Dixon No. 426).

An intermediate stage of the replacement of the francolite is a bright lemon-yellow substance, opaque to transmitted light. It is believed to be ferrous iron-stained francolite or perhaps even an iron phosphate.

CRINOID STEMS AND HEMATITE

Although it is not possible to identify the bulk of the pieces of crinoid stems, they constitute much of the fossil fragments of these ores. In cross section they are seen to contain a tubular network probably of chitin, and show all the stages of replacement by hematite, of the lake-red variety. These replaced crinoid stems, water-worn, form the nuclei for oölites composed of hematite and chamosite in thin coatings. This outer zone may consist of several hundred layers which fill the hollows in the water-worn surfaces of the crinoid stem fragments (Pl. 1, D–K).

CHEMICAL RECASTING

GENERAL DISCUSSION

Inspired by Diller's (1898, p. 138) and Smyth's (1892, p. 488) experiments, the author repeated the treatment of oölitic grains with hydrochloric acid. The acid removed the hematite and left shells of amorphous silica.

TABLE 6.—*Chemical analyses of Clinton ores**

No.	1	2	3	4	5	6	7	8
Ore Bed	Furnaceville	Sterling Sta.	Verona	Verona	Wolcott F	Wolcott F	Wolcott F	Westmoreland
Locality	Ontario	Sterling Sta.	Verona	Verona	Wolcott	Wolcott	Wolcott	Oneida Co.
Analyst	Cummings	Touceda	Touceda	Touceda	Touceda	Brakes	—	—
Source	—	—	—	—	—	—	Beck	Chester
SiO_2	11.74	5.13	9.69	7.06	8.56	23.98	6.0	13.09
TiO_2	—	—	.244	tr	—	.225	—	—
Al_2O_3	.48	4.99	3.21	11.13	5.04	7.26	7.5	5.99
Fe_2O_3	51.85†	48.7	48.71	40.92	44.38	44.71	51.5	64.11
MnO	tr	.003	tr	tr	tr	—	—	.19
CaO	7.34	12.6	13.8	14.68	13.71	9.15	24.5***	5.85
MgO	3.76	7.75	4.23	3.84	7.37	2.92	7.75†††	2.69
S	.028	.096**	.141**	.025**	.072**	.987**	.03†	.31
P_2O_5	.494††	.477	2.38	1.02	1.58	.549	.5†	.53
CO_2	21.46†	18.35	15.45	16.3	18.8	9.6	—	6.08
H_2O	2.76†	1.904†	2.33	4.88	.488†	.26	2.75	1.45
Total	99.912	100.000	100.185	99.855	100.000	99.641	100.53	100.29

No.	9	10	11	12	13	14	15
Ore Bed	Westmoreland	Westmoreland	Westmoreland	Westmoreland	Westmoreland	Westmoreland	Kirkland
Locality	Cayuga Co.	Franklin	Franklin	F & C Co.	Franklin Co.	Average	F & C Co.
Analyst	Cummings	Smyth	Britton	Britton	Britton	Chester	Sullivan
Source	—	Smock	—	Putnam	Putnam	Putnam	Eckel
SiO_2	6.01	11.57	29.72	9.98	11.34	12.63	8.71
TiO_2	—	—	—	—	—	—	—
Al_2O_3	.95	3.92	4.13	2.4	3.91	5.45	3.67
Fe_2O_3	49.97	69.17	42.97	79.98	71.82	63.0	30.24
MnO	.47	.19	.37	tr	1.63	.15	—
CaO	13.96	5.8	8.57	1.54	3.97	6.2	20.64
MgO	7.8	2.27	1.96	.3	2.21	2.77	7.84
S	.11**	.28	.837	nil	nil	.23	.15
P_2O_5	.8	1.726	1.534	1.239	2.096	1.5	.75
CO_2	19.39	4.75	6.72†	4.39	2.00†	6.15	24.78
H_2O	.45	.324†	2.75†	.171	1.024†	2.77	3.22†
Total	99.91	100.000	99.561	100.000	100.000	100.85	100.00

* Taken from Newland and Hartnagel (1908).
† Assumed by the author.
** SO_3.
†† P.
*** $CaCO_3$.
††† $MgCO_3$.

Following the lead of A. O. Hayes (1915), the author assembled chemical analyses of Clinton iron ores from New York, from Newland and Hartnagel (1908), and recast 15 of them, calculating the norms. This was not an easy task because it was necessary

to devise a set of "rules and regulations" for these sedimentary rocks. The lack of the determination of the alkalis prevented calculation of the small amounts of feldspar and the potash-bearing clay minerals. Nonseparation of ferrous iron from ferric iron reduced the accuracy of the amount of siderite.

By making a number of necessary assumptions reasonable results were obtained. Whenever possible a check was made with modal analyses from thin sections of the same bed. In some cases the norms were recalculated several times until an approximate check with the modes was secured (Fig. 6A).

COMPOSITION OF THE OÖLITIC HEMATITE

Studies of the slides of the Clinton ores reveal two distinct colored iron-bearing minerals; the Tuscan-red material is the onion-layered oölitic chamositic hematite, while the lake-red matter is nearly pure hematite, as shown by chemical tests. Microscopic analyses show that these two minerals occur in about equal proportions, varying from 30 to 68 per cent chamositic-bearing hematite. The norms calculated from chemical analyses show that the chamosite content varies from 3 to 33 per cent of the amount of the chamositic hematite. The average is close to 15 per cent. Combining these two figures, the modal chamositic hematite contains about 30 per cent normative chamosite. This amount apparently varies, which would account for the variation in the colors as observed under reflected light.

TYPES OF ORES

There are three types of ore. Unfortunately the miners' terms—"fossil" and "oölitic"—do not make the adequate distinction now possible. One ore, chiefly in the west, hence called the western ore type, is characterized by a higher carbonate and consequently lower hematite content. The carbonate is partly fossil fragments which form the nuclei of the oölites. The chamositic hematite coatings are oölitic although the shapes are much more irregular than similar structures in the eastern portion of the State.

The other type, centering around Clinton, is also oölitic but is more spherical and hence is known as "oölitic ore". This implies that the western type is not oölitic, which is not the case. The eastern type has less carbonate, hence fewer fossil fragments and, consequently more hematite. The quartz, on the other hand, is higher. In other respects the quantitative amounts of the other minerals are about the same.

The Kirkland, the "red flux" of the miners, is essentially a limestone with hematite of low chamosite content (Pl. 3L).

There are three types of ore, two of which are oölitic (Fig. 6B). The sizes of the oölitic types are shown in Figure 7.

ROUNDNESS OF QUARTZ NUCLEI

The quartz centers of the oölites of the eastern type of ores are well rounded (Pl. 3, B-J). They appear to have been rounded by chemical attack by the iron-bearing solutions. Microscopic measurements, however, show that both the roundness (Krumbein, 1941) and the circularity (Rittenhouse, 1943) are functions of size.

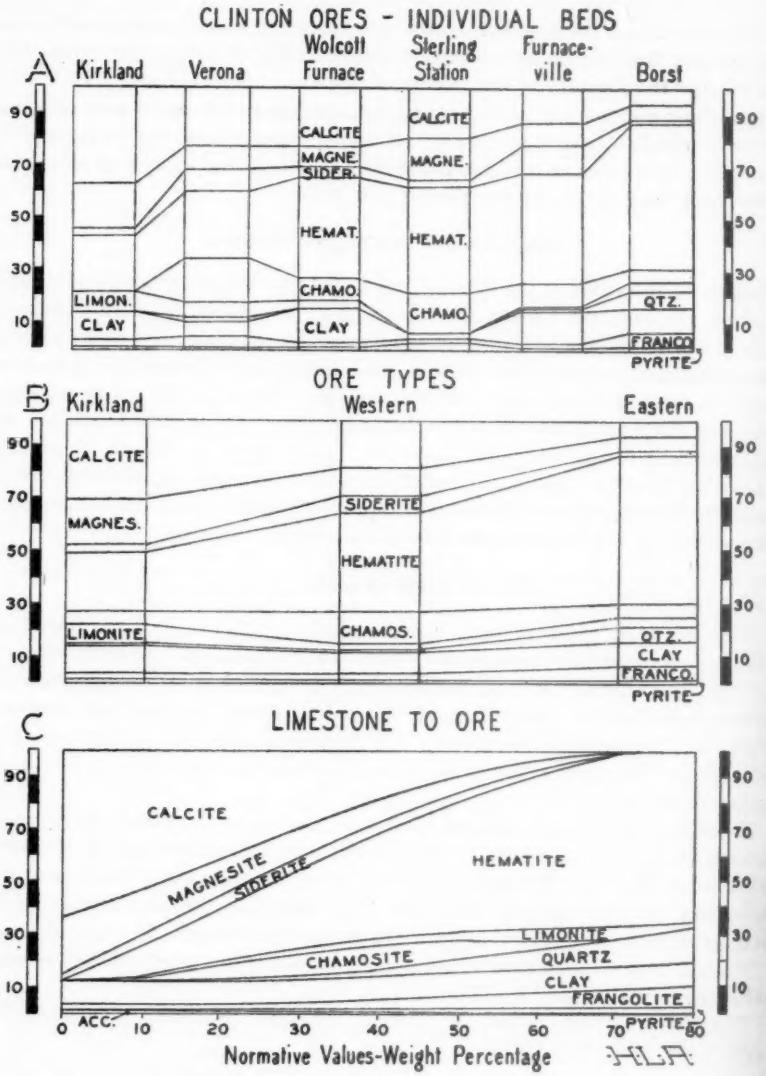


FIGURE 6.—*Normative composition of Clinton ores*

Based upon chemical analyses collected by Newland. Diagram shows changes in the formation of the ore from original limestone. For "Borst", last column in A, read "Westmoreland."

The small grains or groundmass quartz grains in these oölitic ores join and continue the curves of the quartz nuclei (Fig. 8). In other words, the roundness of the quartz

nuclei is just what is expected for grains of that size. This suggests that the rounding was due to mechanical wear and not to attack by the iron-bearing solutions. This observation suggests that the detrital quartz grains of nuclear size were well rounded before the iron solutions were introduced. In Figure 8 the size of the

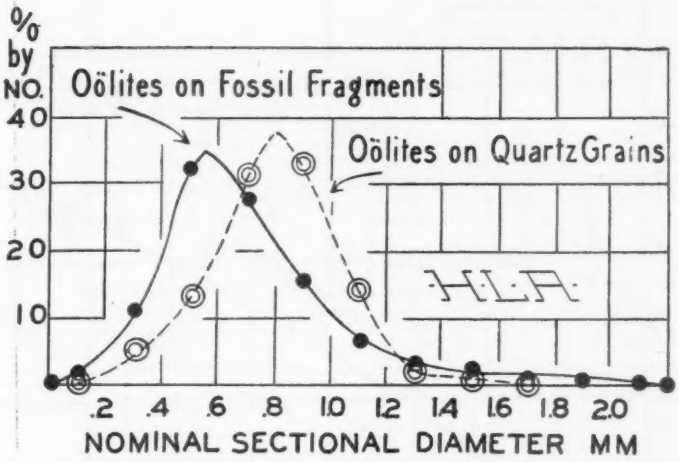


FIGURE 7.—*Frequency-distribution curves of the two types of oölites*

Showing size, as nominal sectional diameter (Wadell, 1935), by per cent by number. The mode size of those on quartz grains is 0.8 mm., while that built on fossil fragments is 0.55 mm. Based upon 841 separate measurements.

nominal sectional diameter (Wadell, 1935) is shown in millimeters on a logarithmic base; the grade scale is metric (Alling, 1943).

QUANTITATIVE CHANGES IN FORMING THE ORE

The 15 norms of Clinton ores make it possible to calculate the composition of the original rock which, in the presence of the iron-bearing solutions by interstitial filling and replacement, became the ore. The quality and quantity of the data are below desirable standards; nevertheless the results obtained are illuminating. These are offered with the knowledge that they are somewhat speculative.

In comparing the norms of these ores, it is readily observed that they are all similar. This is remarkable since they come from widely scattered areas within the State and from different ore horizons.

The most striking variation is that as the hematite increases, the percentage of the normative calcite decreases. The amount and composition of the carbonate changes with the content of the hematite in the ore. These normative minerals are not independent variables, but dependent.

The normative analyses reveal small irregularities which seem relatively insignificant because of the date of the chemical analyses, difficulties in recasting, and the natural variation of these iron ores. To reduce these variations, analyses of the same

ore beds were averaged, the averages plotted, and the resulting curves smoothed. Thus a single diagram can show the major changes that probably occurred in the transformation from the original limestone to the richest ore (Fig. 6C).

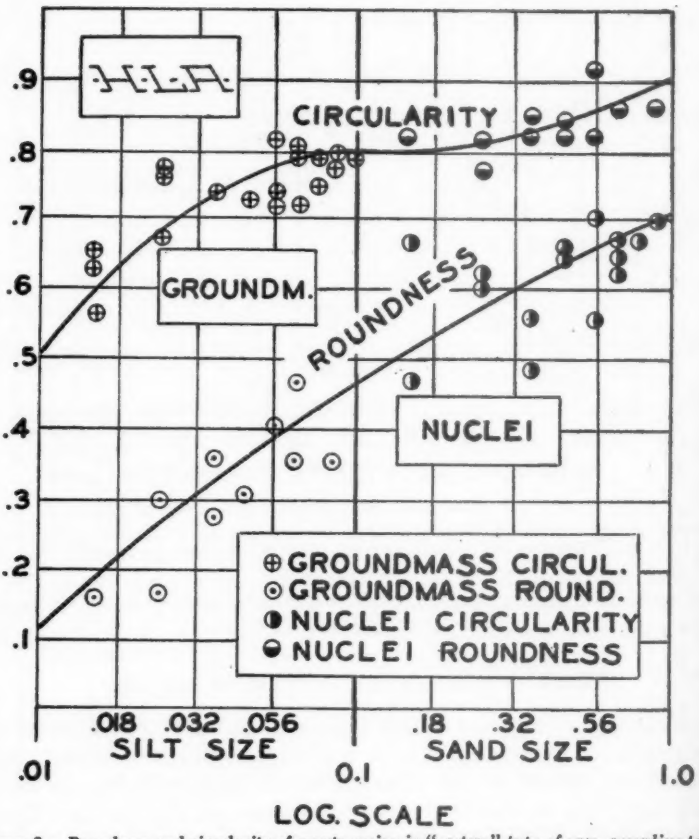


FIGURE 8.—*Roundness and circularity of quartz grains in "eastern" type of ores, according to size*

Based upon visual methods of Krumbein (1941) and Rittenhouse (1943). The size is nominal sectional diameter. The grade scale is metric (Alling, 1943). The roundness of the quartz grains is continuous from the small angular grains of the 'groundmass' to the larger, well-rounded grains that compose the centers of the oölites. This suggests that the roundness, for size, is normal for waterworn clastics. (Lower two symbols, nuclei, reversed.)

In addition to the increase of hematite towards the right, towards the eastern type of ore, Figure 6 shows increase of the pyrite, the francolite, and, above all, the quartz. The quartz represents a stratigraphic change in the composition of the rock, independent of the replacement by hematite. The increase in the phosphate may be entirely due to stratigraphic causes, but it may represent in part the concentration of francolite derived from replaced limestone.

SOURCE OF THE IRON

It has been assumed that the iron was derived by weathering of older rocks at some distance from these shallow basins and was transported by rivers. No details are available of course. Actually our knowledge of the chemistry involved is exceedingly scant (Van Bemmelen, 1909). We do not know in what form the iron was and, if different from what it is today, how and what changes took place. It is obvious that sufficient iron must have been available. It may have been precipitated by contact with sea water (Spring, 1899) as "limonite" and oxidized and dehydrated later. Rutherford (1936) commented that

> "the origin of the sedimentary iron ores in some of the marine formations of the Paleozoic. . .is a problem not yet satisfactorily explained." (p. 1213). In the. . .Silurian. . .we see a recurrence of the concentration of iron in the seawater. . .The conception of a change in the amounts and ratio of the soluble substances in seawater throughout geological time is not contrary to the general doctrine of uniformatarianism." (p. 1214)

There may have been more iron in the Clinton seas than is present today.

Perhaps the iron was derived from laterite. That would require relatively high temperatures with low rainfall (Lang, 1920) and would imply a tropical or subtropical climate with good underground drainage. A. O. Hayes (1915, p. 71) agrees with this concept. Such conditions favor the removal of the silica and a residue rich in iron and aluminum hydroxides. Such iron compounds could well be colloidal provided they were stabilized by protective organic acids. Recently Roy (1945) has shown that much of the silica in ground waters can be in true solution.

A. O. Hayes (1915, p. 75) has suggested that iron in the form of iron "salts of organic acids, chlorides, sulphates, ferrobicarbonates, etc., would sink to the bottom ...[where] ammonia [would be] formed ... from decaying organic matter [which] might produce, directly or indirectly, a precipitate of iron oxide." Hayes applied this theory to the Wabana ores of Newfoundland, where there is a general absence of limestone. In the Clinton ores which are dominantly calcareous, especially in western New York, the carbonates may have assisted in the precipitation process. It was not necessary, however, as the Wabana ores show.

The author shares with Quirke (1943, p. 665) the thought: "I wonder if the presence of the limestone is in any way essential to the deposition of hydrous and anhydrous ferric oxide?" He was applying this question to the Steep Rock Lake deposits near Atikokan, Ontario.

The iron-bearing solutions were more than solutions of iron as is demonstrated by the importance of the presence and function of the chamosite. They must have contained both silica and alumina. The microscopic appearance of the thin skins of chamosite suggests colloidal conditions. It is significant that most of the aluminosilicates are later than the lake-red hematite which lacks or is low in chamosite. It is logical that with time the solutions from the source areas changed in composition.

It has been suggested that the iron was precipitated as chamosite or some other variety of iron-bearing chlorite which was broken down to some form of iron oxide (Penrose, 1892; Spurr, 1894a). A. O. Hayes (1929) commented that the iron content of chamosite is not as high as that of hematite and that there is not enough silica

nor alumina now in the ores to satisfy the formula for chamosite. It is unlikely that these substances have been leached out since there is no evidence of such activity.

Chamosite without any appreciable associated hematite is found in the Clinton section (slides loaned by N. C. Dale) (Pl. 3 G, H, J.). A few oölites from the upper bed at Clinton show a lake-red hematite coating on top of Tuscan-colored oölitic chamositic hematite. This indicates that the two types varied in some localities from the usual order.

DIAGENESIS OF CLINTON IRON ORES

The Clinton rocks were deposited in very shallow seas, some of which may have been interior basins, possibly with abnormally high salinity. The shore lines were constantly lifting as Sanford (1935; 1936) and Gillette (1940; 1947) pointed out. In these seas existed abundant marine life, now found as fossils which show considerable rounding, the result of wave wear. With the introduction of the iron-bearing waters, these iron compounds infiltrated into the pore space between the detrital grains and into the fossil cavities vacated by the organisms upon death. There the iron, perhaps as "limonite", first stained the francolite, chitin and the carbonates and then gradually replaced them. Later the introduced solutions contained more silica and alumina and deposited chamosite in addition to the potential hematite. Evidently this was colloidal and formed very thin, alternating onion-skin, concentric-layered coatings upon the early, relatively chamosite-free iron compounds.

Meanwhile there was considerable recrystallization of the carbonates, especially those low in magnesium. Some silica cemented the detrital grains not already bound by the iron compounds; some became chert, and some of that is now quartz. The replaced carbonate may have assisted in the reorganization of the remaining carbonates, resulting in additional deposition upon the margins of the recrystallized calcite.

Upon burial these deposits became more compact and dehydrated and oxidized, changing the original iron compounds to hematite. The carbonates further recrystallized.

ORIGIN OF CLINTON ORES

A reasonably thorough, though not necessarily intensive, survey of the extensive literature has convinced the author that much of the disagreement about these ores is due to (1) the fact that some theories are based upon a limited set of observations, with consideration of only a portion of the great variety of conditions, and (2), the lack of agreement on the *meaning* of the nomenclature used.

There are essentially two theories: one that the ores are *primary*, and the other that they are due to *replacement*. It is necessary to examine the meaning of these two terms. "Primary" means first. It suggests that something occurred later. "Primary" implies it is part of the original processes that formed these sedimentary rocks.

"Replacement", on the other hand, connotes those processes by means of which a mineral has been replaced or substituted by another. The word itself does not

specify *when* the process occurred. There is no time element expressed nor necessarily implied. "Primary" is a *time* term. "Replacement" is a *kind* of geologic process. Using "primary" and "replacement" in the above manner, the author concludes that both terms are appliable to the Clinton ores. There are many ore deposits of which this is true.

Extensive studies of large collections of thin sections leave no doubt in the author's mind that the processes that produced the ores were many and occurred over some little time: all were part of the diagenesis of the rock. They are "primary" ores in that sense, but replacements were part of the ore-forming processes nevertheless.

The following was kept in mind in studying the Clinton ores: (1) They are of the bedded type, in the strictest sense. (2) They are thin, long lenses, which pinch out and come in again. (3) They are very extensive. (4) They are associated with sediments of shallow-water origin. (5) Both the so-called "fossil" and "oölitic" ores were formed by the same process. (6) Solutions carrying iron were introduced into the shallow seas and there precipitated by reaction with the carbonates, the marine salts, and possibly by bacteria and oxidation. (7) They are integral members of a stratified series. (8) They are not the result of replacement *long after* the deposition and lithification of the rocks; otherwise the ores would be "pockety", and the iron would stain adjacent rocks. (9) Ground waters played no essential part in the formation of the ore. The conclusion is that the ores are the result of "*diagenetic replacement.*"

Newland and Hartnagel (1908, p. 50) followed Smyth (1894) and said "that the theory of sedimentary origin is fully substantiated . . . [and] is the only explanation at all compatible with the conditions." The uniformity of character of these ore beds "is consistent only with a sedimentary derivation" (p. 51.) This is true along the outcrop as well as in depth. "Enrichment by solution and deposition of the iron has not occurred in the New York beds."

The author, however, does not place so much emphasis on the theory that part of the iron had been deposited as a carbonate just because small amounts of siderite were found. He assumes that the normative siderite is a part of the normal carbonate of the original limestone. Furthermore, it is thought that the siderite is less soluble than either calcite or magnesite and, consequently, has been concentrated with the removal of the replaced calcite. Such a theory, however, is not necessary for the Wabana ores, and one may question its application to the Clinton ores. A little of the siderite may well be the result of a reaction between calcium and magnesium carbonates and the iron solutions. Some small amount appears to be later as an introduced mineral.

PARAGENESIS

Cayeux (1909, p. 284–285), in his explanation of the Clinton ores, proposed an elaborated series of changes and replacements. Neither A. O. Hayes (1915) nor the author is able to determine the basis for many of these suggestions. Hayes is particularly positive that many lack justification. But unquestionably Cayeux was ahead of his time and saw, as some of us are only beginning to appreciate now, the complexity of the diagenesis of sediments.

There are three kinds of changes that appear to have taken place:

Fillings, such as those of the zooecial tubes of bryozoans, the ossicles and canals of crinoids, and interstitial spaces between mineral grains.

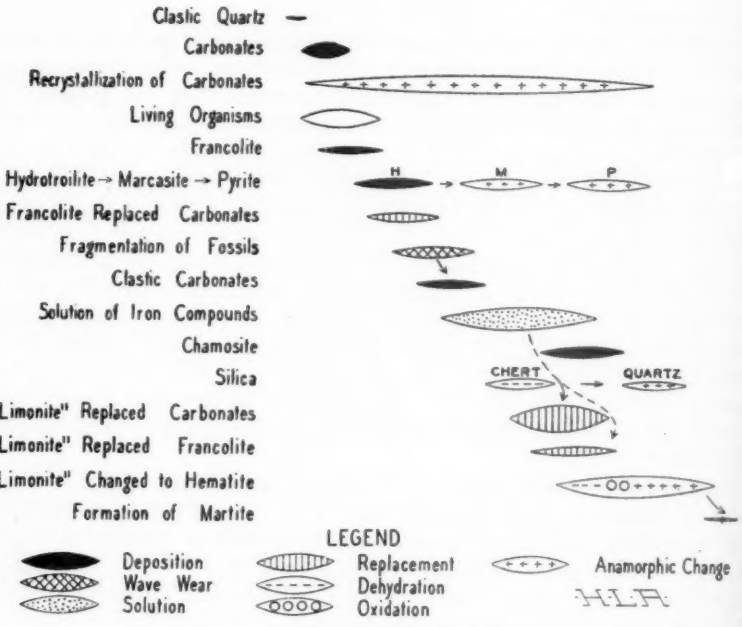


FIGURE 9.—*Paragenesis of Clinton iron ores*

Qualitative diagram indicating order of events with passage of time. Some portions are rather speculative; many replacements are omitted due to lack of space.

Stainings, of many mineral substances by various compounds of both ferrous and ferric iron.

Replacements, probably after first staining, chiefly of limestone.

These processes, of course, frequently overlap.

In the lean ores, where the hematite does not completely obscure details, the infiltration of clay minerals, clastic quartz grains, and clastic grains of carbonate fill the vacated zooecial tubes and ossicles of crinoids. The phosphatic substances, believed to be chiefly francolite, are very common.

Except for the clastic quartz, which seems to be neutral in all of these changes, iron stains affected these minerals in all degrees. Because of the yellows and browns it is believed that some of this iron is still ferrous. There is, therefore, much justification for suggesting that perhaps all of it was ferrous in the beginning.

The following replacements have been noted: (1) Collophane (francolite) replaced fossil fragments. (2) Collophane (francolite) replaced clastic calcite. (3) Iron-stained francolite, both yellow and brown varieties, replaced fossil fragments. (4)

Lake-red hematite replaced fossil fragments. (5) Lake-red hematite replaced francolite, first by staining. (6) Pyrite replaced francolite. (7) Chert replaced carbonates, especially fossil fragments. (8) Chert replaced lake-red hematite. (9 Late siderite replaced lake-red hematite.

The following recrystallizations and anamorphic changes have been noted: (1) Fossil calcite to recrystallized calcite. Some of this may be primary. (2) Some clastic clacite to recrystallized calcite. Only a moderate amount. (3) Ferrous sulphides, such as hydrotroilite, changed to marcasite. (4) Marcasite changed to pyrite. (5) Hematite changed to martite.

Figure 9 attempts to show some of these events in the proper order. Most of these are regarded as details of diagenesis. While the diagram is illuminating, it should not be regarded as too positive since much of it is speculative.

CONCLUSIONS

Petrologic studies of the Clinton ores reveal that they are "primary" ores due to the diagenesis of marine limestones in the presence of solutions well supplied with iron compounds. The formation of ore was accomplished by infiltration, staining, and replacement–diagenetic replacement. If the original iron compounds were ferrous, most of the iron has become oxidized to the ferric condition and largely dehydrated.

Toward the end of the period of the introduction of the iron compounds, siliceous material became available and formed chamosite which together with the hematite formed oölitic structures in thin onion-skin layers. If the nuclei were fossil fragments, the "western" type of ore resulted; if the nuclei were sand-size (two-dimensional measurements) quartz grains, the "eastern" type was formed. The Clinton seas may have been richer in iron than modern oceans.

If the introduced solutions had precipitated carbonates or silica instead of iron compounds, much less study would have been given to the resulting rocks. The implications of this paper appear to be that similar processes prevailed in the diagenesis of the more common kinds of marine sedimentary rocks. Such rocks will receive more attention in the future.

REFERENCES CITED

Alling, H. L. (1943) *A metric grade scale for sedimentary rocks*, Jour. Geol., vol. 51, p. 259–269.

——— (1945) *The use of microlithologies as illustrated by some New York sedimentary rocks*, Geol. Soc. Am., Bull., vol. 56, p. 737–756.

——— (1946) *Quantitative petrology of the Genesee Gorge sediments*, Rochester Acad. Sci., Pr., vol. 9, p. 5–63.

Bradley, W. H. (1946) *Coprolites from Bridger formation of Wyoming*, Am. Jour. Sci., vol. 244, p. 215–239.

Cayeux, Lucien (1909) *Les minerios de fer oölitkique de France*, Fosc. Minerois der Fer Primires, Paris Impro. Nat., p. 284–285.

Chadwick, G. H. (1918) *The stratigraphy of the New York Clinton*, Geol. Soc. Am., Bull., vol. 29, p. 327–368.

Clarke, J. M., and Ruedemann, R. (1903) *Guelph fauna in the State of New York*, N. Y. State Mus., Mem. 5.

Diller, J. S. (1898) *The educational series of rock specimens collected and distributed by the U. S. Geol. Survey*, U. S. Geol. Survey Bull. 150, p. 138.

Foerste, A. S. (1891) *On the Clinton oölitic iron ores*, Am. Jour. Sci., 3d ser., vol. 41, p. 28–29.

Gillette, Tracy (1940) *Geology of the Clyde and Sodus Bay quadrangles, New York*, N. Y. State Mus., Bull. 320, p. 5–179.

——— (1947) *The Clinton of western and central New York*, N. Y. State Mus., Bull. 341, 191 p.

Hall, James (1843) *Natural history of New York, Part 4, vol. 3 Geology*, p. 60. N. Y. State Geol. Surv., Albany, N. Y.

Hayes, A. O. (1915) *Wabana iron ore of Newfoundland*, Geol. Survey Canada, Mem. 78, 163 p.

——— (1929) *Further studies of the origin of the Wabana iron ore of Newfoundland*, Econ. Geol., vol. 24, p. 1–4.

——— (1931) *Structural geology of the Conception Bay region, and of the Wabana ore deposits of Newfoundland*, Econ. Geol., vol. 26, p. 44–64.

Hayes, C. W. (1894) *Description of the Sewanee sheet (Tenn.)*, U. S. Geol. Survey, Atlas, Chattanooga Folio, no. 8.

Kemp, J. F. (1895) *Ore deposits of the United States* (2nd ed.), Scientific Publishing Co. New York, p. 104.

Kimball, J. P. (1891) *Genesis of iron ores by isomorphous and pseudomorphous replacement of limestone*, Am. Geol., vol. 8, p. 352.

Krumbein, C. W. (1941) *Measurements and geological significance of shape and roundness of sedimentary particles*, Jour. Sedim. Petrol., Fig. 1, p. 68.

Lang, R. (1920) *Verwitterung und Bodenbildung als Einführung in der Bodenkende*, Stuttgart, p. 119.

Loomis, F. B. (1903) *The dwarf fauna of the pyrite layer at the horizon of the Tully limestone in Western New York*,, N. Y. State Mus., Bull. 69, p. 892.

McCallie, S. W. (1908) *Fossil iron ores of Georgia*, Ga. Geol. Survey, Bull. 17, p. 185–194.

McGlammery, Winnie (1931) *The dwarf fauna of the Furnaceville iron ore at Rochester, New York*, Unpublished manuscript, Univ. Rochester.

Newberry, J. S. (1880) *The genesis of our iron ores*, School Mines Quart. 2, p. 13–14.

Newland, D. H., and Hartnagel, C. A. (1908) *Iron ores of the Clinton formation in New York State*, N. Y. State Mus., Bull. 123, 76 p.

Penrose, R. A. F., Jr. (1892) *The Tertiary iron ores of Arkansas and Texas*, Geol. Soc. Am., Bull. 3, p. 44.

Quirke, T. T. (1943) *Hydrothermal replacements in deep seated ore deposits of the Lake Superior region*, Econ. Geol., vol. 38, p. 662–666.

Rittenhouse, G. (1943) *A visual method of estimating two-dimensional sphericity*, Jour. Sedim. Petrol., vol. 13, Fig. 1, p. 80–81.

Roy, C. J. (1945) *Silica in natural waters*, Am. Jour. Sci., vol. 243, p. 393–403.

Rutherford, R. L. (1936) *Geologic age of potash deposits*, Geol. Soc. Am., Bull, vol. 47, p. 1213–1214.

Sanford, J. T. (1933) *Stratigraphy of Niagara Gorge*, 16th Inter. Geol. Cong., Guidebook A-4, p. 99.

——— (1935) *The "Clinton" in Western New York*, Jour. Geol., vol. 43, p. 169–183.

——— (1936) *The Clinton in New York*, Jour. Geol., vol. 44, p. 797–814.

Smyth, C. H., Jr. (1892) *On the Clinton iron ore*, Am. Jour. Sci., 3d ser., vol. 43, p. 487–496.

——— (1894) *Die Hämatite von Clinton in den östlichen Vereinten Staaten*, Zeits. prak. Geol., p. 304–313.

——— (1911) *The Clinton type of iron-ore deposits*, In *Types of ore deposits* (ed. by H. F. Bain), Mining and Scientific Press, San Francisco, p. 33–52.

——— (1918) *On the genetic significance of ferrous silicate associated with the Clinton iron ore*, N. Y. State, Mus., Bull. 207–208, p. 175–198.

Spring, W. (1899) *Ueber die eisenhaltigen Forbstoffe sedimentaerer Erdboden und ucher den Wahrscheinlichen Uhrsprung fer rotten Felsen*, Neues Jahrb. Mineral., vol. 1, p. 47–62.

Spurr, J. E. (1894a) *The iron-bearing rocks of the Mesabi range*, Am. Geol., vol. 13, p. 335.

——— (1894b) *The iron-bearing rocks of the Mesabi range in Minnesota*, Geol. Nat. Hist. Survey Minn., Bull. 10.

Stose, George W. (1924) *Origin of Clinton hematite ores*, Econ. Geol., vol. 19, p. 405–411.

Ulrich, E. O., and Bassler, R. S. (1922) *American Silurian formations*, Md. Geol. Survey, Silurian vol.

Van Bemmelen, J. H. (1909) *Ueber das Vorkommen, die Zusammensetzung, und die Bildung von Eisenhaufungen in und unter Mooren*, Zeit. anorg. Chemie, vol. 22, p. 313–379.

Wadell, Hakon (1935) *Volume, shape and roundness of quartz particles*, Jour. Geol., vol. 43, p. 250–280.

Wesenberg, and Medd, C. (1901) Dansk. Geol. for. 7, p. 79–87, 159.

White, Donald E. (1947) *Diagenetic origin of chert lenses in limestone at Soyatal, State of Queretaro, Mexico.*, Am. Jour. Sci., vol. 245, p. 49–55.

REFERENCES TO CHEMICAL ANALYSES

Beck, L. C. (1842) *Mineralogy of New York*, W. and A. White and J. Visscher, Albany, N. Y.

Chester, Albert H. (1881) *The iron region of Central New York*, Address delivered before the Mercantile and Manufacturing Association, Utica, New York.

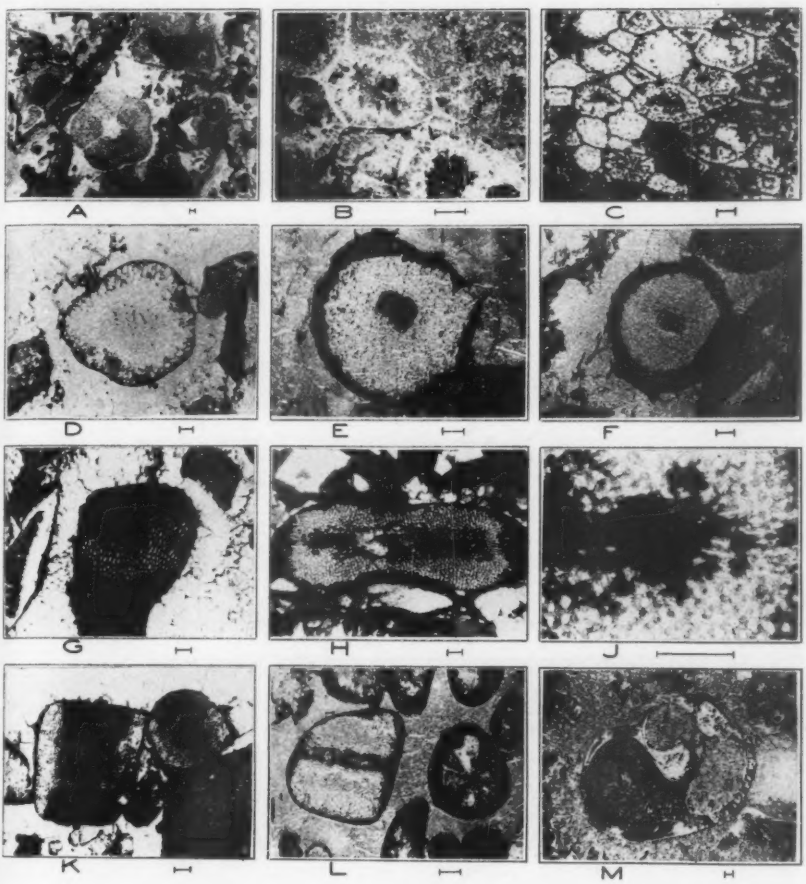
Eckel, E. C. (1905) *The Clinton hematite*, Eng. Min. Jour., vol. 79.

Putnam, B. T. (1886) *Notes on the samples of iron ores collected in New York*, 10th Census of U. S. vol. 15.

Smock, John C. (1889) *Report on the iron mines and iron ore districts in the State of New York*, N. Y. State Mus., Bull. 7.

University of Rochester, Rochester, N. Y.

Manuscript Received by the Secretary of the Society, February 14, 1947.



FOSSIL FRAGMENTS IN CLINTON ORES

(Scale line: 0.1 mm.)

A. Crinoid and bryozoan fragments in nonhematitic Reynales limestone, Furnaceville horizon, Lockport quadrangle. Slide courtesy of J. T. Sanford. Transmitted light, ×6.6.

B. Possibly a coral in Furnaceville, Rochester. Reflected light, ×41.5.

C. Possibly a coral in Furnaceville, Densmore Creek. Transmitted light, ×32.

D. Crinoid section showing chitinous tubular network, partially replaced by hematite, Densmore Creek. Transmitted light, ×23.

E. Transverse cross section of crinoid stem with central canal filled with hematite, the margin coated by chamositic hematite. Furnaceville ore, Rochester. Reflected light with a mirror beneath the slide. ×32.

F. Similar to E, Furnaceville ore from Densmore Creek. ×32.

G. Crinoid, showing network largely replaced by hematite (black), Furnaceville ore, Densmore Creek. Transmitted light, ×23.

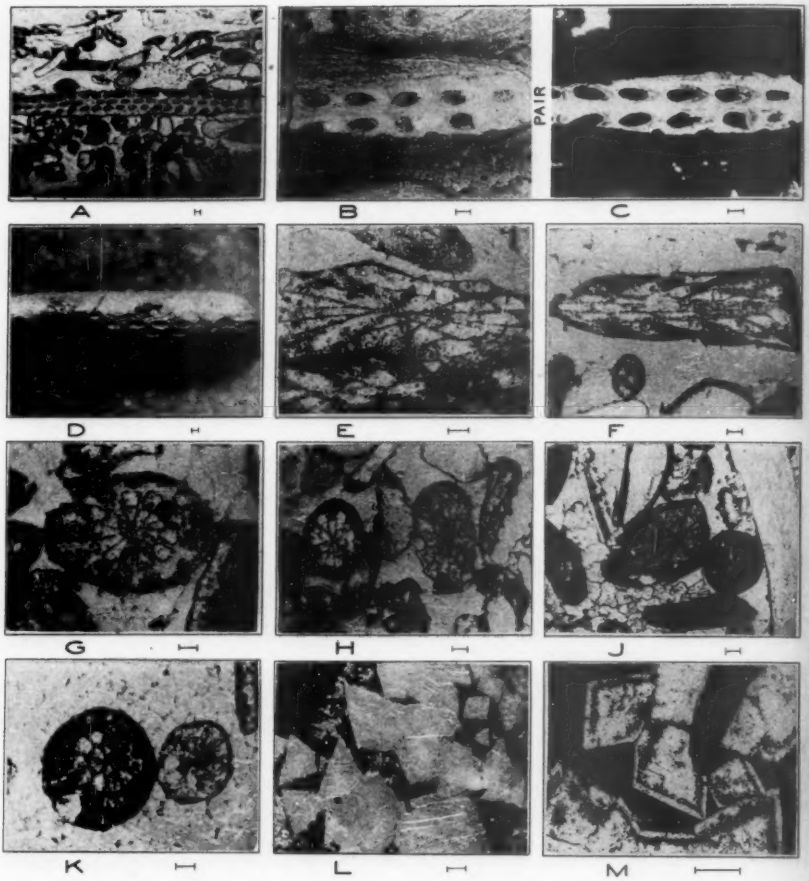
H. Tubular network of crinoid fragment partially replaced by hematite and margin coated by chamositic hematite. One inch oölitic layer, Clinton, N. Y. Specimen courtesy of Virginia Hoyt Jones. Transmitted light, ×23.

J. Portion of above, higher magnification, ×114.

K. Diagonally sliced crinoid stem showing network and canal filled by hematite, and the margin coated by chamositic hematite. Furnaceville ore, Rochester. Transmitted light, ×23.

L. Rounded, longitudinal section of crinoid stem, with canal filled by hematite and the margin coated by chamositic hematite, Furnaceville ore, Rochester. Reflected light with a mirror beneath slide. ×32.

M. Chamositic hematite oölite in a possible gastropod, Kirkland "ore," "red flux," Clinton, N. Y. Specimen courtesy of Virginia Hoyt Jones. Reflected and transmitted light, ×12.



THIN SECTIONS

A–K, *Helopora fragilis*, Scale line, 0.1 mm.

A. *H. fragilis* — longitudinal section, apertures filled by hematite and coated by chamositic hematite, Furnaceville ore, Densmore Creek. Transmitted light, ×6.5.

B. and C. A pair. B — reflected light, C — transmitted light. Furnaceville ore, Ontario, N. Y. ×23.

D. *H. fragilis* from the McGlammery collection, Furnaceville ore, Rochester, showing coating of chamositic hematite. ×9.5.

E. Diagonally sliced section, partially replaced by hematite, Furnaceville ore, Densmore Creek. Reflected light with a mirror beneath the slide. ×32.

F. Similar to E. Same locality, same lighting. ×23.

G. Cross section, similar to E and F. Both transmitted and reflected light. ×32.

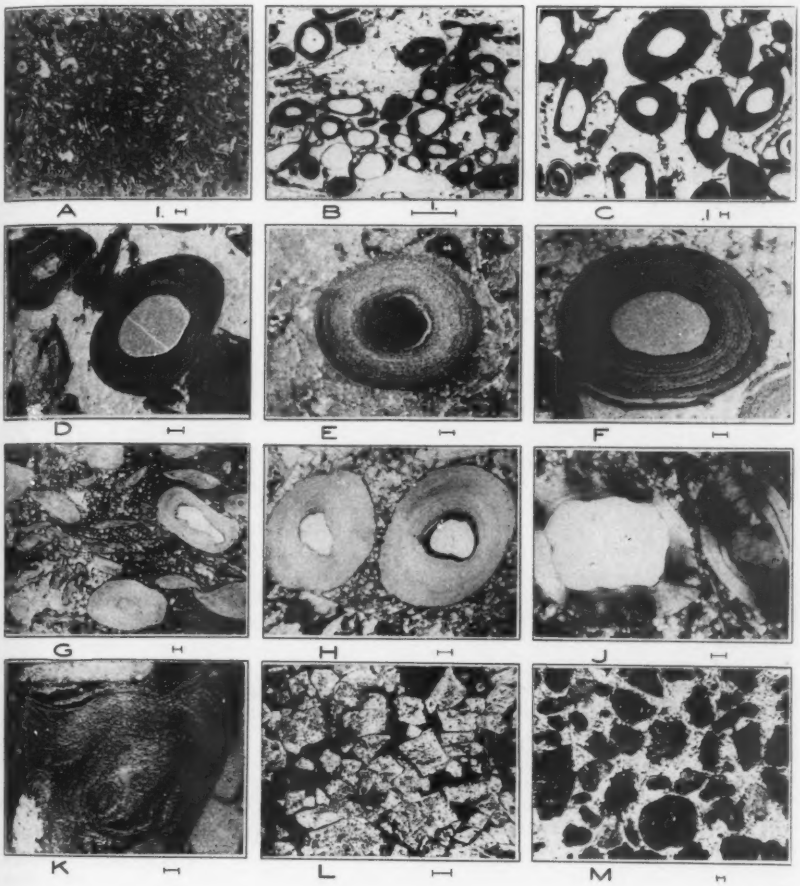
H. Cross sections, Furnaceville ore, Rochester. Reflected light with a mirror beneath slide. ×23.

J. Similar to H, transmitted light only. ×23.

K. Two cross sections, partially replaced by hematite. Reflected light with a mirror beneath the slide.

L. Hematite (black) in wedge-shaped grains between recrystallized carbonate, Reynales limestone, just above the Furnaceville, Densmore Creek. Specimen courtesy of Bernard H. Dollen. Reflected light with a mirror beneath the slide. ×32.

M. Recrystallized carbonate rhombs with authigenous margins, Lower oölite bed, Clinton, N. Y. Specimen courtesy of Virginia Hoyt Jones. Transmitted light, ×66.



THIN SECTIONS

(Scale line: 0.1 mm., except in A and B where it equals 1.0 mm.)

A. Fossil fragments, all partially replaced by hematite and coated with chamositic hematite, Furnaceville ore, Densmore Creek. Entire slide transmitted light, ×1.5.

B. Many chamositic hematite oölites with quartz centers, Franklin mine, upper part of bed, Clinton, N. Y. Specimen courtesy of C. A. Hartnagel and J. G. Broughton. Transmitted light, ×6.5.

C. Very similar to B. ×12.

D. Very similar to B and C, except reflected light with a mirror beneath slide. ×23.

E. Similar to D. Both reflected and transmitted light. ×23.

F. Oölite with quartz center, lower oölitic bed, Clinton, N. Y. Specimen courtesy of Virginia Hoyt Jones. Reflected light with a mirror beneath slide. ×23.

G. Oölites of chamosite, chamosite layer, Clinton, N. Y. Slide loaned through courtesy of N. C. Dale. Transmitted light, ×12.

H. Two oölites of chamosite with quartz centers. Kindness of N. C. Dale. Transmitted light, ×23.

J. Similar to H, but with cross nicols showing polarization crosses. Slide loaned by N. C. Dale. ×23.

K. Disrupted chamositic hematite oölite, showing separation of layers, Furnaceville ore, Ontario, N. Y. Reflected light with a mirror beneath slide. ×23.

L. Wedge-shaped grains of hematite between rhombs of recrystallized carbonate. Kirkland ore, Clinton, N. Y. Specimen courtesy of Virginia Hoyt Jones. Transmitted light, ×32.

M. Oneida conglomerate, Utica quadrangle. Quartz grains (black), some well rounded, cemented by pyrite (white). Specimen courtesy of W. L. Grossman. Reflected light, ×12.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 58, PP. 1019-1084, 4 PLS., 27 FIGS. NOVEMBER 1947

ARCHEAN GEOLOGY OF THE TURKU DISTRICT IN SOUTHWESTERN FINLAND

BY ANNA HIETANEN

CONTENTS

	Page
Abstract	1020
Introduction	1021
The rocks and their classification	1023
The oldest complex	1023
Kinzigites	1023
Amphibolites and leptites	1027
Intrusive rocks	1032
Classification	1032
Intrusives of the first cycle	1033
Trondhjemite series	1033
Charnockite series	1035
Pöytyä diorite and associated rocks	1051
Intrusives of the second cycle	1052
Rocks of the granite series	1052
Post-tectonic intrusives	1054
Diabase dikes	1054
Anorthosite	1054
Rapakivi	1056
Quartz porphyry dike	1056
Chemical and mineralogical composition of the rocks of the charnockite and granite series	1057
Comparison with the other regions and the origin of the charnockite	1062
Structures	1065
Structure of the sedimentary and volcanic rocks	1065
Structure of the syntectonic intrusives of the first cycle	1067
Structure of the intrusives of the second cycle	1068
Rocks derived by the intense recrystallization of the effusive rocks and the leptites	1070
Migmatization	1071
Granitization	1072
Kakola granite	1073
Summary and conclusions	1078
References cited	1081

ILLUSTRATIONS

Figure	Page
1. Location of the Kalanti and Turku districts in southwestern Finland	1021
2. Geologic map of the southwestern contact zone of the Pöytyä diorite area	1022
3. Derivation of biotite at expense of cordierite	1025
4. Cordierite kinzigite, 3 km southwest of the Merimasku church	1026
5. Relict agglomeratic structure in a fine-grained layer of the volcanic complex	1029
6. Hypersthene-bearing layers in kinzigite in the island Pahaluoto, Rymättylä	1032
7. Unaltered diopside surrounded by hypersthene which changes into cummingtonite	1041
8. Alteration of hypersthene to biotite along the margins and to serpentine along the cracks	1041

9. Alteration of hypersthene to cummingtonite in its margins 1042
10. Hypersthene surrounded by primary diopside and secondary cummingtonite 1043
11. Alteration of hypersthene to biotite in microcline-bearing charnockites 1044
12. Biotite myrmekite around a hypersthene grain 1045
13. Hypersthene altered to serpentine and cummingtonite along the margins and to biotite in the center 1045
14. Alteration of diopside to hornblende 1046
15. Cordierite changing into biotite 1049
16. Garnet-bearing trondhjemite, Luonnonmaa 1050
17. Granodiorite cutting the migmatitic veins discordantly 1054
18. Diopside in lamellae parallel to (010) and dome planes in hypersthene 1054
19. Diagram showing orientation of the optic axes and axial planes of a hypersthene and the lamellae 1055
20. Main oxides of the rocks of charnockite and granite series 1058
21. Niggli values of the rocks of the charnockite and granite series 1060
22. Norms of the rocks of the charnockite and granite series 1061
23. Bedding in the kinzigite in Rusko 1066
24. Inclusions of volcanics in granite-trondhjemite near Lake Hirvijärvi 1068
25. Younger pegmatite vein transects the granite-trondhjemite and the dark inclusions ... 1068
26. Rocks of the microcline granite series penetrating leptites 1069
27. Inclusions of amphibolite in microcline-granite 1069

Plate — Facing page
1. Geologic map of the Turku district (In pocket, end of paper).
2. Photomicrographs 1082
3. Photomicrographs 1083
4. Photomicrographs 1084

ABSTRACT

The Turku area is situated in the islands and along the sea coast in southwestern Finland. On the northwest it joins the Kalanti area which has been mapped and described earlier (Hietanen, 1943). The oldest rocks are kinzigites (cordierite-garnet gneisses) and amphibolites, both highly metamorphosed and migmatized. Kinzigites are of sedimentary origin and show bedding locally. Amphibolites are volcanic or sedimentary as are also the finer-grained leptites which occur with them.

The intrusive rocks are grouped as follows: (1) intrusives of the first cycle, (2) intrusives of the second cycle, and (3) the post-tectonic intrusives.

(1) The oldest intrusives occur as concordant oblong bodies in the oldest rocks. The acidic members are poor in potash, belonging to the trondhjemite series. Intermediate and basic rocks are members of either trondhjemite or charnockite series. Texture is granoblastic. Remnants of older rocks and basic inclusions are common. In small bodies lineation and cleavage parallel the same structures in the country rocks. The cores of the larger plutons are structureless or show a vertical lineation. Many occurrences among charnockitic and trondhjemitic rocks are clearly formed metasomatically or metamorphically. The bulk of these rocks, however, show features common in igneous rocks.

(2) The intrusives of the second cycle (the granite series) are rich in potash and penetrate the older rocks concordantly and also disconcordantly. The structures of the narrow bodies parallel the walls; the centers of the larger ones are structureless.

(3) The youngest intrusives in the area are diabase dikes, anorthosite, and rapakivis.

Migmatization and granitization of the oldest rocks are a result of joint action of long duration of (1) introduction of feldspars and quartz, and (2) metamorphic differ-

entiation. The processes started contemporaneously with the folding and emplacement of the oldest intrusives and continued after the folding until emplacement of pegmatites of the red microcline granite series.

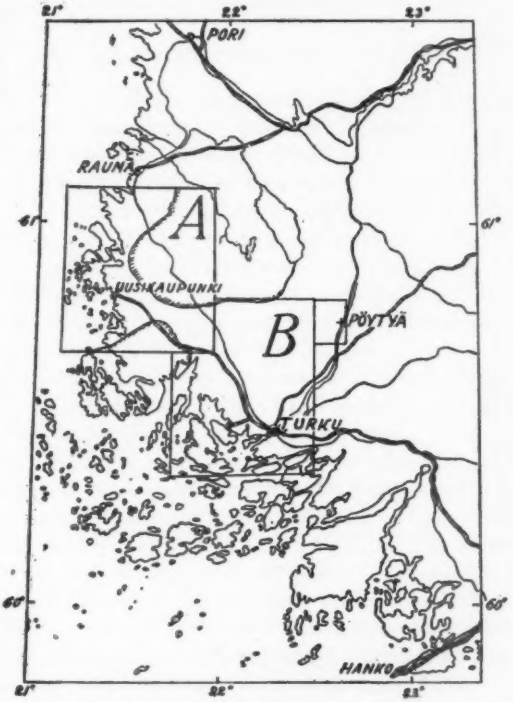


FIGURE 1.—*Location of the Kalanti (A) and Turku (B) districts in southwestern Finland*

INTRODUCTION

The area under discussion is in the coastal area and in the islands around the town of Turku (Åbo) in southwestern Finland (Fig. 1, B). On the north it joins the Kalanti region (Hietanen, 1943) and the areas occupied by the rapakivis of Vehmaa and Laitila (Fig. 1, A). On the south it is bounded by the areas investigated by Hackman (1923), Hausen (1944), and Metzger (1945). Rocks similar in petrography and structures continue from the present area eastward to the parishes of Pöytyä and Yläne (Fig. 2) which are in the western part of the old geologic map of Loimaa (Wilkman, 1898, Sheet no. 32). The trondhjemites and granite-trondhjemites of the eastern part of the present area grade into a more basic differentiate, the diorite of Pöytyä, which occupies vast areas in Pöytyä. The boulder of orbicular diorite from Pöytyä described by Eskola (1938) was found west of this diorite, in the area occupied by the migmatites, whose older parts consist of kinzigites (cordierite-garnet gneiss) intercalated with hornblende-bearing leptites and amphibolites, both probably of volcanic origin. All three are invaded by various members of the Pöytyä diorite series.

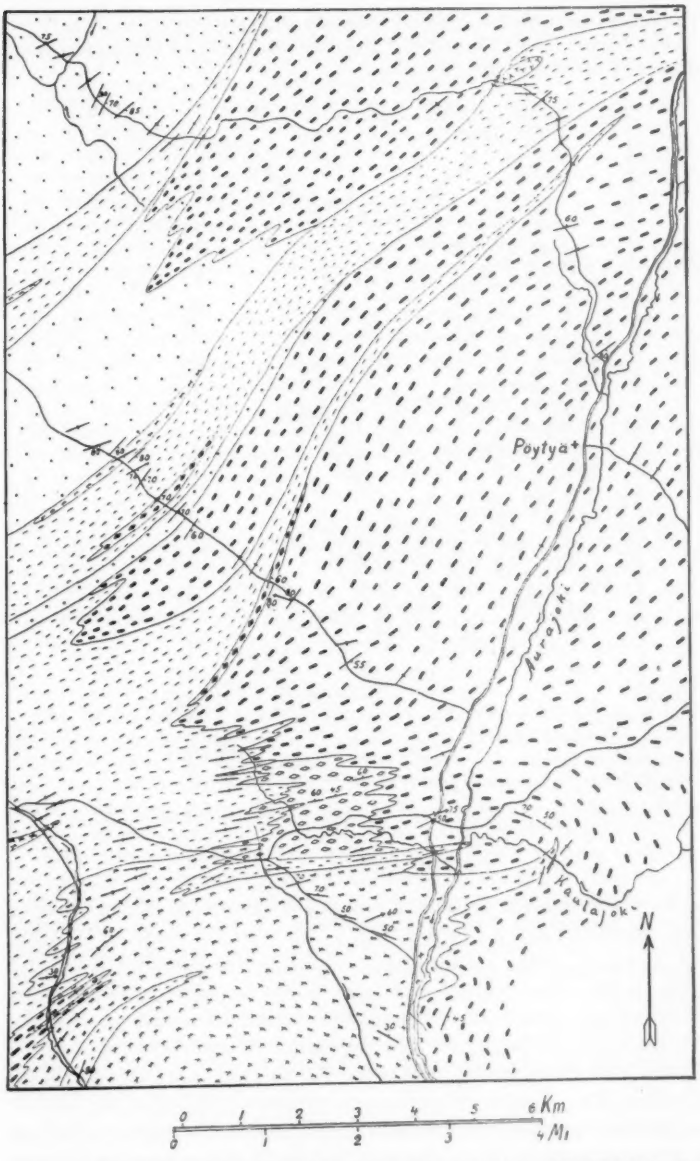


FIGURE 2.—*Geologic map of the southwestern contact zone of the Pöytyä diorite area*
Joins the map of the Turku district. Structural symbols are the same as on Plate 1.

Field investigations for the present work were carried out during the summers 1943–1945 and were supported by the Geological Survey of Finland. I acknowledge

my gratitude to its Director, Professor Dr. Aarne Laitakari. Thanks are also due Mr. O. von Knorring for making the chemical analyses and Mrs. Lyyli Orasmaa for drafting the map, Plate 1.

The specimen numbers in the text refer to the locality numbers on Plate 1.

THE ROCKS AND THEIR CLASSIFICATION

Every rock type of the Kalanti district occurs also in the adjacent Turku region (Pl. 1). There are also several other rock series. The oldest complex which forms the older part of the migmatites consists of kinzigites, amphibolites, and leptites. The kinzigites are generally more metamorphosed and granitized than those in the Kalanti district. Part of the amphibolites and leptites are of a sedimentary origin, and another part volcanic. Both cover wide areas, whereas in Kalanti the amphibolite lenses are rather small, and the true leptites are lacking.

The intrusive rocks may be divided into four separate series: (1) a trondhjemite series, (2) the Pöytyä diorite and associated rocks, (3) the charnockite series, and (4) the red microcline granite and associated basic rocks.

The rocks of the trondhjemite series represent the oldest intrusives in the area and possess all characters of syntectonic type. The Pöytyä diorite and associated basic and acidic rocks may be correlated with the trondhjemites. Some of the rocks of the charnockite series are gneissic, and the contacts are generally concordant. Many larger plutons, however, have an internal structure independent of the structures in the surrounding rocks. These rocks are considered somewhat younger than the trondhjemite series but still syntectonic and thus older than the red microcline granite and associated rocks, which, except for the rapakivi and diabase dikes, form the youngest rock series in the region and are clearly post-tectonic. The red microcline-rich pegmatites, which are abundant everywhere, are considered to represent the residual liquids of these granites. The pegmatitic veins of the migmatites may be artheritic (Sederholm, 1899; 1907) or venitic (Holmquist, 1921) depending on the possible neighborhood of an intrusion, the physico-chemical conditions, and differential movements during the metamorphism. Accordingly their age varies locally, and often there are two series of them in the rock. The garnet- and cordierite-bearing granites in Turku and in its neighborhood, called Kakola granite by Laitakari (1934), contain in addition to palingenetic and venitic material also residual liquids from microcline granite. Its genesis would thus be comparable to that of the vein material.

THE OLDEST COMPLEX

KINZIGITES

The kinzigites of the northern part of the area under discussion are a continuation of the kinzigites of the Kalanti area and are petrologically and structurally very similar to them. Their detailed description is not repeated here. Figures 6 and 12 in Kalanti report (Hietanen, 1943) give a fairly good picture of their structure. The alternation of cordierite- and garnet-bearing layers is considered to be due to the bedding of the original clays. Blue or dark-gray cordierite and garnet ranging from 2 mm. to 2 cm. are present. They may occur in the same layer or in separate ones that range from a few cm to several meters in width.

The kinzigites of the Askainen-Lemu zone are fine- to medium-grained and contain garnet as well as cordierite. In the kinzigite these minerals are usually not megascopically visible, but in the granitic veinlets which are abundant in this zone they are larger and often partly idiomorphic.

No cordierite has been found in the fine-grained, quartz-rich kinzigites of the southern part of the parish Askainen and in wider areas in Masku. These beds resemble the sandy layers of the normal kinzigites. (*Cf.* analyses 8 and 9, Hietanen, 1943.) The feldspar is exclusively soda-lime feldspar and is often richer in anorthite (An_{30-34}) than the plagioclase in the normal kinzigites. The amount of quartz ranges from 40 to 60 per cent, and that of the biotite from 12 to 20 per cent. The structure is granoblastic. A part of the quartz occurs as smaller rounded grains, common also in normal kinzigites (Pl. 2, fig. 2). The biotite scales are well oriented parallel to the schistosity. Magnetite and zircon are common accessories; a few monazite grains are present.

Bedding is visible in the alternation of finer- and coarser-grained layers. The beds are generally 1 or 2 m thick, and some are thicker. In many outcrops the bedding has been obliterated by the advanced migmatization. Its orientation, however, can be determined by the orientation of the cleavage, which parallels the bedding in this area.

Most outcrops, and all outcrops on the highest hills, in the parish of Mietoinen consist of migmatite, whose older part is kinzigite and in which the pegmatitic vein material rich in potash feldspar constitutes from 60 to 90 per cent of the rock. As a rule the migmatites of the small outcrops consist of about 60 per cent kinzigite and about 40 per cent vein material. The veins have penetrated the kinzigite parallel to the only *s*-plane[1] which can be recognized in the rock. They are generally ½ to 2 cm. broad and rather regularly scattered. In most outcrops such a "normal" migmatite is cut by broader pegmatite veins and stocks, which raise the amount of total pegmatite to 80 per cent or more. The pegmatite veins and stocks contain abundant cordierite and garnet which in many high hills are the only evidence that the older rock was a kinzigite. The alignment of these minerals often shows marks of the old structure. Well-preserved kinzigites in this area are found in only a very few small outcrops.

The layers rich in cordierite seem to be most resistant to weathering and migmatization. For example, the cordierite kinzigites south of the Merijärvi charnockite pluton and west and southwest of the Merimasku church are very well preserved. The layers are easy to recognize in the field because of their rough-weathering surface. Also some layers north and south of the volcanic and leptitic rocks northeast of Mynämäki contain only cordierite. The cordierites ranging from 1 to 2 cm. in diameter are clearly visible in the weathering surface as darker and deeper-weathered patches. The intercalated sandy layers are about 20 cm. thick and fine-grained.

The texture of the kinzigites is like that in the kinzigites of Kalanti. Some thin sections show clearly a reaction between potash feldspar, biotite, and cordierite similar to that described by Parras (1941) from the Lohja region. In the cordierite-garnet

[1] *S*-plane or *s*-surface is used as defined earlier by Sander (1911, p. 286). It is a planar structure, which may be a result of stratification, cleavage, or mechanical slip.

kinzigite no. 1216, the biotite is derived at the expense of cordierite. The sillimanite needles formed by the alumina released in this reaction are included in the cordierites near the contact with biotite. Also small fresh biotite needles occur in the same zone (Fig. 3). Thin sections, no. 933 and 1115 (Pl. 2, fig. 1), show another type of reaction

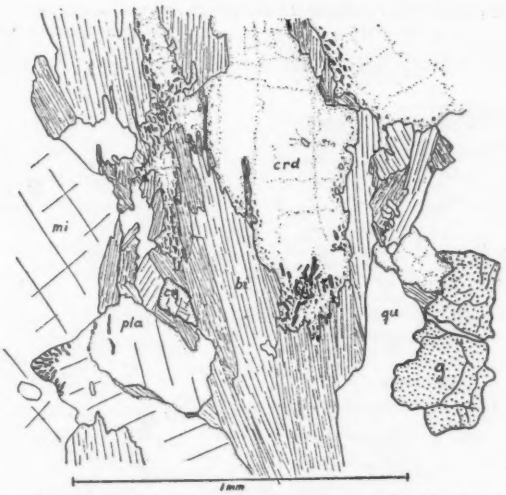


FIGURE 3.—*Derivation of biotite at expense of cordierite*

Small biotite scales and sillimanite needles occur in the cordierite near the contact with biotite. Cyanite prism which is included in biotite was also formed at the expense of the alumina released in the reaction. crd = cordierite, bi = biotite, cy = cyanite, g = garnet, mi = microcline, pla = plagioclase, qu = quartz. Cordierite kinzigite 2½ km. south-southeast of Lake Savojärvi, Pöytyä (no. 1216, Pl. 1).

zone between biotite and cordierite. In no. 933 the cordierite changes first to pinite, then there follows a zone of light-grayish-green chlorite, and around the grain biotite. The reaction: Biotite + sillimanite + quartz ⇄ cordierite + potash feldspar is reversible. Quartz set free in the reaction occurs as worm-shaped inclusions in biotite. Some of the cordierite metablasts with sillimanite needles have a margin with abundant quartz and biotite inclusions. The writer believes this margin and the cordierites with quartz and biotite inclusions (crd. II in Fig. 4) are younger than the cordierites with sillimanite needles (crd. I in Fig. 4). The microcline metablasts contain quartz and biotite inclusions in a manner similar to the younger cordierite (Pl. 2, fig. 2) and have obviously been formed during the same period of recrystallization. Thus the course of metamorphism has here been largely the same as in the Kalanti area. First the kinzigites were metamorphosed by moderate heat in the amphibolite facies. The main minerals at this stage were: quartz, plagioclase, cordierite, and biotite. The heat then rose and caused an advanced metamorphism which attacked part of the biotite. Cordierites of the second generation and the microcline metablasts were formed at the expense of that biotite. The reaction cordierite → biotite as shown in Figure 3 would mean a local retrogressive metamorphism. Reaction in no. 1115 occurs in connection with granitization. The refrac-

tive index of the biotite is: $\gamma = 1.643 \pm 0.001$. According to Winchell's tables its composition is: $4H_2O \cdot 2K_2O \cdot 11(Fe, Mg)O \cdot 3Al_2O_3 \cdot 11SiO_2$. This biotite has reacted with the small amount of sillimanite present in the kinzigite of the first stage. The reaction may be expressed as follows:

$$\underset{\text{2 biotite +}}{2 \times 4H_2O \cdot 2K_2O \cdot 11(Fe, Mg)O \cdot 3Al_2O_3 \cdot 11SiO_2} + \underset{\text{20 sillimanite +}}{20 \times Al_2O_3 \cdot SiO_2} + \underset{\text{37 quartz}}{37SiO_2} \rightleftarrows$$
$$\underset{\text{11 cordierite +}}{11 \times 2(Fe, Mg)O \cdot 2Al_2O_3 \cdot 5SiO_2} + \underset{\text{4 microcline +}}{4 \times K_2O \cdot Al_2O_3 \cdot 6SiO_2} + \underset{\text{8 water}}{8H_2O}$$

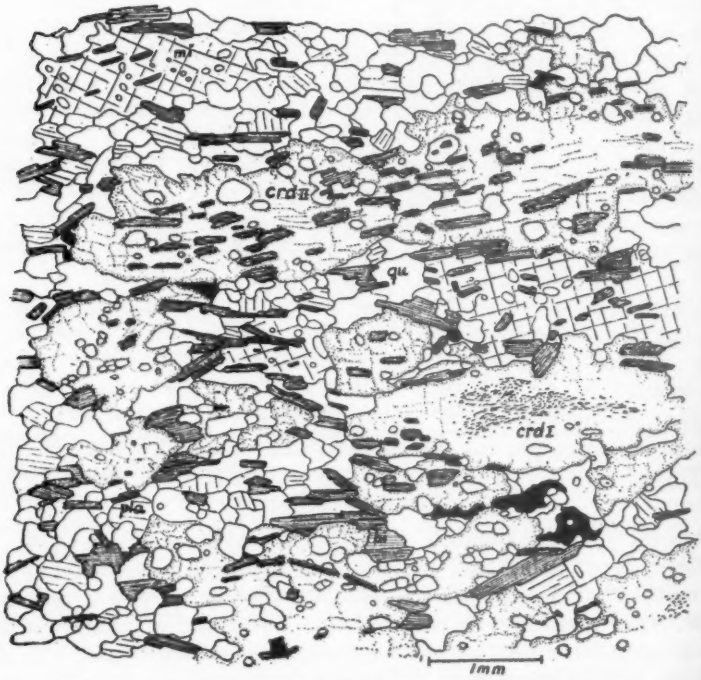


FIGURE 4.—*Cordierite kinzigite, 3 km southwest of the Merimasku church* (No. 892, Pl. 1). Later recrystallized cordierite (crd II) and microcline occur as larger grains, including quartz (qu) and biotite (bi).

Muscovite does not occur in these rocks. The boundaries between cordierite and biotite in Merimasku are sharp and differ thus from the boundaries described above. Apparently equilibrium has been attained in the Merimasku area. The mineral assemblage cordierite-microcline-biotite is critical of the facies and implies a higher temperature than the amphibolite facies. Because sillimanite is needed for the formation of cordierite at the expense of biotite, the excess of alumina available in the kinzigites of the amphibolite facies has an important role. The sillimanite included in cordierite metablasts did not take part in the reaction, and thus equilibrium was attained even though the chemical analyses still show an excess of alumina. (Cf. analysis 7, Hietanen, 1943.)

AMPHIBOLITES AND LEPTITES

The well-exposed outcrops of amphibolites and fine-grained dark leptites in the northern part of the area show clearly traces of the structures of lava flows and tuffs. Well-preserved relict structures were found on both sides of the small road leading to the farm Lahomaa south of the village Vehmalainen, Karjala. Medium- and fine-grained amphibolite and leptite layers alternate with uralite porphyrites and plagioclase porphyrites. The approximate thickness of the various layers from north to south in these outcrops is as follows:

	Meters	
	>4	leptite
X	3	coarse-grained, dark uralite porphyrite
	4	plagioclase porphyrite
	4	uralite-plagioclase porphyrite
IX	2	leptite
	1	fine-grained uralite porphyrite
VIII	2	leptite
	3	medium-grained uralite porphyrite
VII	3	fine-grained leptite
	2	plagioclase porphyrite
	36	fine-grained leptite
	2	coarse-grained light-colored leptite (sedimentary)
VI	6	fine-grained leptite
	10	plagioclase porphyrite
V	15	fine-grained leptite
	25	medium-grained uralite porphyrite
IV	33	fine-grained leptite with magnetite amygdules
	3	plagioclase porphyrite
	5	medium-grained uralite porphyrite
III	1	fine-grained leptite, uppermost part containing magnetite amygdules
	1½	plagioclase porphyrite
	21	medium-grained uralite porphyrite
II	24	fine-grained leptite, folded thickness about 10 m.
	1	stretched, medium-grained uralite porphyrite
I	5	fine-grained leptite
	>15	coarse-grained uralite porphyrite

Each uralite porphyrite layer grades into fine-grained leptite to the north, and many of the northernmost beds of the leptite contain small amygdules. It seems therefore that each layer of uralite porphyrite and leptite represents a separate ancient lava flow, the bottom of which would be to the south. The structure of the formation supports this view (Pl. 1). The volcanic rocks occur as a large lenticular body intercalated and folded with the kinzigites. The outcrops south of Vehmalainen are on the northern limb of a large anticline, whose axis dips eastward. Thus the bottom of the formation must lie on the south side of the northern limb.

The coarse-grained uralite porphyrites consist of hornblende ($\gamma = 1.659 \pm 0.001$ measured in the spec. no. 970) with some plagioclase. Sphene and magnetite are common accessories. Some hornblende occurs as grains 5–10 mm in diameter or groups of grains of medium size. The boundaries between these groups and larger grains often still show crystallographic faces of augite. The groundmass is fine-grained and granoblastic. In many localities the large hornblende crystals form most

of the rock—*e.g.*, nos. 927, 956, 970. The minerals show random orientation. The rock is black with glittering crystal surfaces of hornblende and has therefore been quarried for tombstones—*e.g.*, in locality no. 970. Generally this variety occurs as lenticular bodies in the medium-grained stretched uralite porphyrite.

In the medium-grained variety the minerals are the same as in the coarse-grained, but there is more plagioclase. It is therefore lighter in color and usually has well-developed cleavage and lineation.

The plagioclase porphyrite layers are characterized by light-colored "phenocrysts" or plagioclase, 1 to 2 cm in length, in the dark, fine-grained groundmass. Microscopic examination shows complete recrystallization of the groundmass and "phenocrysts"; both are granoblastic. The structure of these beds as seen in the field is thus very similar to that of the common porphyrites. Under the microscope, however, it is not easy to recognize their origin because of the granulation of the phenocrysts. The fine-grained groundmass consists chiefly of quartz, plagioclase (An_{40}), and hornblende; biotite, magnetite, and sphene are accessories.

Layers containing both plagioclase and uralite phenocrysts occur, for example, on the west side of the main road south of Vehmalainen. In the field the dark uralites and light plagioclases are distinctly visible in a dark-brownish, fine-grained groundmass. Under the microscope these "phenocrysts" also appear to have been granulated (Pl. 2, fig. 3). Feldspar is partly sericitized, and epidote is a common accessory. Many magnetite grains are surrounded by epidote.

Plagioclase, hornblende, and quartz are the main constituents of the fine-grained leptites. Biotite may also occur; magnetite, sphene, and apatite are common accessories. Most of these layers are even-grained and lepidoblastic. Some of them contain large hornblende holoblasts which include numerous rounded quartz and plagioclase grains and also biotite flakes. In some upper layers there are numerous small amygdules: round grains of iron ore surrounded by light-colored minerals. Microscopic examination shows that the ore grains are rimmed by small grains of sphene, and this by plagioclase and quartz (Pl. 2, fig. 4). These plagioclase and quartz grains are larger than those in the groundmass. The amygdules apparently have been vesicles in the upper parts of the lava flows. The light minerals first crystallized along the walls of the small cavities, and with rising temperature the ore filled the center. The titanium content of the ore gave rise to the formation of sphene.

Some leptite layers show a distinct bedding; dark layers, 1 mm to 5 cm thick alternate with the lighter reddish layers 5 mm to 2 cm thick. The dark layers consist of quartz, plagioclase, hornblende, biotite, and some microcline. Quartz and plagioclase form the reddish layers. The amount of microcline varies. These rocks occur as long, narrow lenses between the lava flows and were probably tuffs or weathering products of tuffs and lavas.

Agglomeratic structure was found in some layers of the fine-grained leptite about 1 km south of Lahomaa where the following section was measured:

Meters	
	coarse-grained uralite porphyrite
2	fine-grained, stretched uralite porphyrite
30	leptite

Meters	
15	uralite porphyrite, the middle part less stretched
50	leptite
>5	agglomerate

The agglomerate layer is thus at the base of the formation. The ancient volcanic bombs are darker narrow long lenses in the fine-grained leptitic rock (Fig. 5). The

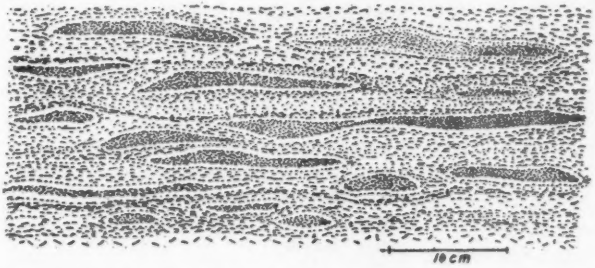


FIGURE 5.—*Relict agglomeratic structure in a fine-grained layer of the volcanic complex* South of Vehmalainen (No. 935, Pl. 1). The dark lenticles are ancient volcanic bombs.

bombs consist of quartz, plagioclase, diopside, and some hornblende. The dark minerals of the groundmass are biotite with some diopside. The minor constituents are the same as in the leptites.

Some leptites differ mineralogically and often also texturally from the leptites described above (sedimentary leptite on Pl. 1). They contain, in addition to quartz and plagioclase, abundant microcline and biotite, but no hornblende. The accessories are magnetite, sphene, apatite, and zircon. These leptites occur as long lenses intercalated with the volcanic rocks. The two largest lenses, 2 and 3 km northeast of Mynämäki, consist of a coarse-grained rock in which a part of quartz and plagioclase occurs in small pebbles. This gives to the rock a sedimentary appearance. The long narrow lens north of Lake Valkamajärvi consists of a fine-grained reddish rock, with distinct bedding. The beds are 1 to 5 cm thick, and their composition varies. Some beds contain only quartz and plagioclase, others have abundant microcline also. The biotite occurs only in certain layers (Pl. 2, fig. 5). The mineral assemblage and structure seem to indicate that these leptites are sedimentary, most likely arkosic. Their special feature seems to be that they occur with the volcanics.

The leptite in a small island west of Rymättylä (no. 1277) contains about 50 per cent microcline and only a little plagioclase. Some small scales of biotite occur as a dark constituent.

Kinzigite layers are also often intercalated with volcanics, especially in the border zones of the formations. This is shown very nicely on Plate 1, near the village of Tarvainen, 6 km northeast of Mynämäki. About 1 km to the south the following rock series occurs:

Meters	
10	pegmatite
50	cordierite-garnet kinzigite
100	leptite

	Meters	
	15	garnet kinzigite
	20	leptite
	20	garnet kinzigite
		leptite containing hornblende
N		garnet kinzigite
↑		leptite with amphibole
|	20	garnet kinzigite
S		fine-grained leptite
		leptite
		garnet kinzigite
	7	uralite porphyrite
	5	garnet kinzigite with abundant garnet
	2	leptite
		coarse-grained uralite porphyrite
		Kakola granite with abundant garnet
		garnet kinzigite
	4	leptite
		garnet kinzigite
		leptite
		garnet kinzigite

Field examination shows that the alternation of the volcanic and sedimentary layers is not due to folding but must be primarily caused by the alternation of clays, tuffs, and lavas. Several lava flows would explain the alternation, and the thickness of separate kinzigite layers may permit an estimate of the length of time between eruptions.

The above features, or traces of them, are visible also in the other amphibolite and leptite areas, and the origin of these rocks is therefore considered to be volcanic. Alternation of coarse-grained uralite or plagioclase porphyrites and fine-grained leptites is common and recognizable in Raisio although the rocks here are more highly metamorphosed than in Karjala. Also the granitization and migmatization have in many localities obliterated the old structure as described below.

Amphibolites of the western part of the area occur as long narrow lenses in the kinzigites parallel to the bedding. Their structure and minerals are similar to those in the Kalanti area. Well-oriented hornblende is the main constituent, and plagioclase is always present. Diopside occurs generally in narrow layers or lenticles parallel to the foliation. These lenticles are 1 to 10 cm long and $\frac{1}{2}$ to $1\frac{1}{2}$ cm. thick. The occurrence and mineral composition suggest a sedimentary origin.

The amphibolites south of Turku are similar to those described by Metzger (1945) from Parainen. In the Turku area the occurrences are small, and no limestones were found. Some diopside amphibolites, however, contain considerable calcite scattered as small grains among the other minerals. On the north shore of the island of Pitkäluoto abundant wollastonite with $\gamma = 1.632 \pm 0.001$, $\beta = 1.630 \pm 0.001$, $\alpha = 1.617 \pm 0.001$ occurs together with pyroxene, garnet, and calcite (no. 1330 on Pl. 1). The garnet has the refractive index $n = 1.770 \pm 0.001$ (grossularite, with about 22 per cent andradite according to Winchell's diagram).

The indices of refraction of the pyroxene $\alpha = 1.716 \pm 0.001$, $\beta = 1.720 \pm 0.001$, $\gamma = 1.736 \pm 0.001$ indicate according to Winchell 66 per cent hedenbergite. Clearly these amphibolites were originally marly clays.

A small lenticle of cummingtonite-garnet amphibolite occurs on the northern part of the small island Tammisluoto, Rymättylä (no. 1277, Pl. 1). Cummingtonite forms larger porphyroblasts and shows polysynthetic twinning (Pl. 2, fig. 6).

Pyroxene amphibolites occur as long narrow lenses in the kinzigites. They are brownish gray and fine-grained and resemble the fine-grained schistose norites which occur as similar small lenticles and are associated with the charnockites. The pyroxene amphibolites contain some hypersthene in addition to abundant diopside and hornblende. If hornblende is the dominant dark constituent, the rock looks like normal amphibolite. The amphibolites in Littoinen east of Turku and in the islands south of Ruotsalainen (no. 1299, Pl. 1) belong to this type. Pyroxenes and hornblende occur often in separate layers or in small lenticular masses. The hypersthene-bearing diopside amphibolites in the eastern part of Hirvensalo (no. 1270) and south of the Rymättylä church (no. 1325) contain abundant diopside and less hornblende. About 5 km from Turku, toward Piikkiö, on the south side of the road an outcrop of diopside-plagioclase rock contains also some hypersthene. A similar rock occurs as a narrow lenticular body in the kinzigite on the north side of the same road about 6 km southeast of Turku.

The structure and mode of occurrence of the pyroxene amphibolites suggest sedimentary origin. They would thus be comparable to the amphibolites south of Turku and those in Parainen (Metzger, 1945) except that a certain amount of hypersthene had crystallized in the pyroxene amphibolites. This indicates metamorphism in the pyroxene-hornfels facies. Some of the small lenticular norite occurrences may also be sedimentary. All these hypersthene-bearing rocks are well distributed within the same area as the charnockites (Pl. 1), *viz.*, south of the zone Masku-Merimasku. The kinzigites in that area near Turku are mineralogically similar to the kinzigites of the northern part of the present district but differ structurally from them. Near Turku these rocks are coarse-grained and generally do not show any bedding. One might suppose that the intense recrystallization has obliterated the bedding which is clearly recognizable farther north. However, the bedding is still clearly visible west of Turku, where hypersthene-bearing rocks are also common. In the island of Ruotsalainen for instance a narrow dark noritic layer occurs in the kinzigite, which shows a distinct bedding and no signs of higher metamorphism than the kinzigites of the cordierite-microcline subfacies. Figure 6 shows similar beds from Pahaluoto, Rymättylä. (*See also* Figure 2 of Plate 3.) The lack of primary structures in the kinzigites near Turku is therefore not necessarily due to higher temperature during metamorphism. It seems rather to be related to a more intense granitization. The abundant cordierite-garnet granites, which are the granitization products of the kinzigites, occur just in this area (Pl. 1).

The occurrence of the foliated, fine-grained gabbros is similar to that of the amphibolites in the kinzigites. As a rule they contain no diopside, and their structure differs somewhat from that of the true amphibolites: the hornblende is less well oriented and occurs as thicker crystals. Some well-foliated rocks rich in hornblende contain no diopside but otherwise appear more like the amphibolites than the gabbros. These rocks are mapped in the color of the amphibolites.

Small occurrences of scarn are common in the small islands west of Rymättylä in

association with the amphibolites and amphibole-bearing leptites. The main minerals are hornblende and diopside, which commonly occur in separate lenticles. Andradite-hedenbergite scarn is also found. The ores, magnetite and pyrrhotite, are scarce.

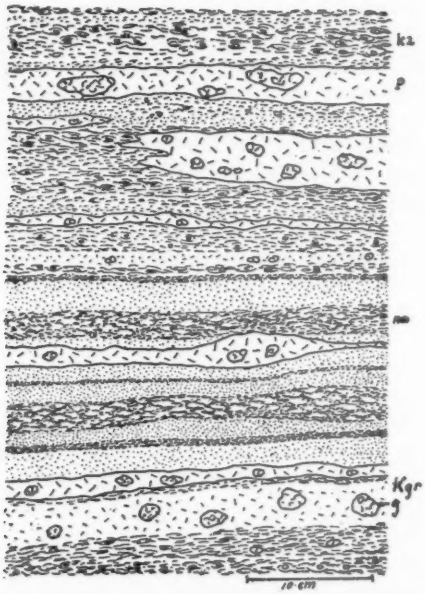


FIGURE 6.—*Hypersthene-bearing layers (no) in kinzigite (kz) in the island Pahaluoto, Rymättylä* (No. 1294, Pl. 1). Pegmatite (P) and granite (Kgr) veins contain garnet (g).

The scapolite occurrence of Laurinkari west of Hirvensalo has been described by Borgström (1915).

INTRUSIVE ROCKS

CLASSIFICATION

The older intrusives may be divided into four petrographically distinct differentiation series. Rocks of the trondhjemite series prevail in the northern part of the area, while the intrusives of the southern part contain usually also microcline. Every gradation from the trondhjemites into granites is represented, and the addition of microcline may be primary or secondary. Among the basic rocks hornblendites, gabbros, and diorites similar to those of the trondhjemite series occur, but in addition there are also norites. The greatest variation, however, occurs among the intermediate rocks. The large plutons of the Masku-Merimasku zone are mapped as charnockites. In addition there are diorite-trondhjemites, granodiorites, and all gradations between charnockites and diorite-trondhjemites. The diorite-trondhjemites on the other hand grade into the diorite of Pöytyä east of Masku. In the

southernmost part of the area microcline granites, associated with granodiorites, diorites, and gabbros, form a separate differentiation series which is distinctly younger than the rocks of the charnockite series.

The petrological classification of the intrusives can be tabulated as follows:

a. INTRUSIVES OF THE FIRST CYCLE

I. Trondhjemite series:
   1. Hornblendites and gabbros
   2. Diorites
   3. Diorite-trondhjemites
   4. Trondhjemites
   5. Trondhjemite-pegmatites

II. Pöytyä diorite and associated rocks:
   1. Hornblendites and gabbros
   2. Diorites
   3. Diorite-trondhjemites and granodiorites
   4. Granite-trondhjemites
   5. Granitic pegmatites

III. Charnockite series:
   A. Basic division
      1. Hornblendites, gabbros and basic norites
      2. Diorites and norites
   B. Intermediate division
      1. Normal charnockites
      2. Trondhjemitic charnockites
      3. Hornblende charnockites (hy → bi)
   C. Acidic division
      1. Trondhjemites
      2. Granite-trondhjemites
      3. Trondhjemitic and granitic pegmatites

b. INTRUSIVES OF THE SECOND CYCLE

IV. Granite series:
   1. Gabbros
   2. Diorites
   3. Granodiorites
   4. Granites
   5. Granite-pegmatites

c. THE YOUNG POST-TECTONIC INTRUSIVES

V. Diabase dikes.

VI. Anorthosite.

VII. Rapakivi.

INTRUSIVES OF THE FIRST CYCLE

Trondhjemite series.—In the northern part of the area the rocks belonging to the trondhjemite series are mineralogically and chemically analogous to the corresponding rocks of the Kalanti area and are therefore discussed only briefly in this paper.

Hornblendites occur as small lenses in the kinzigites and as inclusions of various size in the diorites and the intermediate rocks. In addition to hornblende there is a

small amount of magnetite; apatite and sphene are common accessories. The grain size varies from coarse to medium.

Gabbro occurs in small and rather few areas. The common minerals are hornblende, plagioclase, some biotite, and apatite. Gabbros are dark, medium- to coarse-grained. At the border zones the femic minerals are oriented parallel to the contacts.

A coarse- to medium-grained diorite lens continues from the southwestern part of the Kalanti area into the northwestern part of the Turku area. Hornblende and biotite are the dark constituents. Plagioclase and quartz are the light minerals. Magnetite, apatite, and sphene are common accessories.

Fine- to medium-grained trondhjemites occur in one larger and several rather small oblong bodies in the north (Pl. 1). Their contacts with the adjacent kinzigites and volcanic rocks are always concordant. Megascopically, these trondhjemites resemble the normal trondhjemites of the Kalanti region; a soft bluish tint in their light-gray color is characteristic of both, and they differ in this respect from the reddish-gray and gray granites.

The minerals are quartz, plagioclase (An_{19-25}), biotite, and accessories. The microcline may occur in small interstitial grains and as antiperthite in the plagioclase similar to that described from Kalanti (Hietanen, 1943, p. 52).

The trondhjemitic rocks in the northwestern part of Livonsaari are described separately because they differ structurally from the other trondhjemites and form a complete differentiation series from gabbros and diorites to diorite-trondhjemites and trondhjemites. The gabbroic and dioritic members contain large idiomorphic plagioclase crystals. The cores of the intrusions are unoriented having thus more the appearance of an intrusive than the more or less foliated rocks of the trondhjemite series in Kalanti.

The gabbro in small outcrops at the farm house Alitalo in the village of Kauppinen (no. 906) represents the most basic type. It occurs in narrow lenses in the kinzigites and is folded with them and grades over to coarse, light-bluish-gray trondhjemites, in which a part of the feldspar is in bigger crystals resembling the structure of the acidic diorites. The gabbro is composed of plagioclase (An_{73}), biotite, hornblende, a little quartz, and apatite. The plagioclase crystals are 1½–2 cm. long and 2–3 mm. thick, oriented parallel to the foliation near the contact but at random in the center of the lens.

A diorite crops out north of the gabbro at the sea coast. It is dark bluish gray, coarse-grained, and some of the plagioclase (An_{63}) occurs as larger partly idiomorphic crystals, many of which are granulated. The other minerals are quartz in rounded small grains, hornblende, biotite, magnetite, pyrrhotite, apatite, and zircon. Biotite shows a symplectitic intergrowth with quartz.

The composition of the diorite-trondhjemite corresponds to that of the biotite-bearing diorite-trondhjemites of the Kalanti area. In the more acidic types the plagioclase (An_{32-35}) occurs as larger but less euhedral grains. Biotite is usually the only mafic mineral. The same type forms migmatite with the kinzigite occurring in veins 2 to 100 cm. broad. In places (*e.g.*, locality 909) abundant idiomorphic plagioclase crystals occur in the older sedimentary rocks near the contacts of the trondhjemitic rocks. The sodic feldspar has thus here penetrated the rock metasomatically

UNIVERSITY OF CALIFORNIA
LIBRARY

No.	1		2a	
Loc. no. on Pl. 1 ...	1140	1039	1059	
SiO_2	43.42	47.72	50.05	
TiO_2	0.58	1.86	0.37	
Al_2O_3	11.52	13.06	16.41	
Fe_2O_3	3.18	2.86	1.83	
FeO	11.91	10.07	8.70	
MnO	0.41	0.18	0.32	
MgO	12.05	8.22	7.22	
CaO	12.71	10.64	10.62	
Na_2O	0.97	2.63	3.01	
K_2O	0.88	1.51	0.49	
P_2O_5	1.07	0.19	0.04	
ZrO_2	.....	.....	.....	
S	.....	.....	.....	
H_2O+	1.07	1.18	0.77	
H_2O-	0.23	0.30	0.17	
	100.00	100.42	100.00	
quartz	.....	.....	.....	
plagioclase	$An_{19}8.2$	.....	$An_{46}36.86$	A
microcline	.....	.....	.....	
biotite	9.8	.....	1.95	
muscovite	.....	.....	3.05	
chlorite	.....	.....	.....	
hypersthene	.....	.....	2 73	
diopside	.....	.....	3.76	
hornblende	78.1	.....	51.63	
cummingtonite	.....	.....	.....	
actinolite	.....	.....	.....	
serpentine	.....	.....	.....	
magnetite	1.5	.....	0.11	
sphene	.....	.....	.....	
pyrite	.....	.....	.....	
apatite	2.4	.....	0.09	
zircon	.....	.....	.....	
	100.00	.....	100.00	
si	82.3	104.7	.....	
ti	0.82	0.31	.....	
al	12.90	16.88	.....	
fm	58.30	50.41	.....	
c	25.94	25.00	.....	
alk	2.86	7.71	.....	
qz	−29.14	−26.1	.....	−
k	0.37	0.27	.....	
mg	0.58	0.53	.....	
c/fm	0.44	0.50	.....	
qu	.....	.....	.....	
or	5.20	8.93	.....	
ab	8.26	22.29	.....	
an	24.52	19.40	.....	
en	7.97	4.53	.....	
hy	5.10	2.85	.....	
fo	15.43	11.20	.....	
fa	10.81	8.16	.....	
wo	13.23	13.50	.....	
cor	.....	.....	.....	
mt	4.60	4.13	.....	
il	1.10	3.52	.....	
ap	2.48	0.43	.....	
zr	.....	.....	.....	
H_2O	1.30	1.48	.....	
	100.00	100.42	.....	1

1. Hornblendite, Naantali
2, 2a. Norite, Masku
3. Gabbro, Naantali
4, 4a. Norite, Merimasku
5. Norite, Askainen

TABLE 1.—*Chemical and mineralogical composition of the rocks of the char*

Including their Niggli numbers and norms.

	2a	3	4	4a	5	6	7	8	9	9a	10
	1059	1142	877	877	837	833	881	1087	834	834	835
2	50.05	48.92	51.12	52.40	53.37	53.25	63.59	66.40	66.31	65.21	66.6
6	0.37	0.38	1.11	0.31	0.12	0.59	0.51	0.39	0.71	0.27	0.5
6	16.41	15.73	16.81	16.71	21.00	19.05	16.36	16.05	16.02	16.66	15.8
6	1.83	1.91	1.33	1.44	1.00	1.18	0.69	0.66	1.13	0.68	0.9
7	8.70	7.80	8.00	8.30	6.11	6.63	4.28	3.34	3.29	4.08	2.9
8	0.32	0.28	0.13	0.18	0.08	0.12	0.07	0.05	0.04	0.03	0.0
2	7.22	8.35	6.44	6.58	4.12	3.75	2.31	2.01	1.86	2.55	1.5
4	10.62	11.69	9.12	9.45	10.95	7.73	4.48	3.45	3.72	3.88	3.6
3	3.01	2.88	3.46	3.72	2.69	4.93	4.84	5.67	4.94	5.39	4.8
1	0.49	0.54	0.82	0.19	0.09	1.52	1.95	1.11	1.18	0.83	1.8
0	0.04	0.59	0.16	0.08	0.13	0.23	0.16	0.30	0.04	0.05	0.0
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
	.....	.....	.....	0.08	.....	.....	.....	.....	.....	.....	.....
3	0.77	0.74	1.20	0.50	0.16	0.74	0.57	0.43	0.65	0.28	0.8
0	0.17	0.19	0.20	0.14	0.10	0.28	0.19	0.14	0.23	0.09	0.1
2	100.00	100.00	99.90	100.00	100.00	100.00	100.00	100.00	100.12	100.00	99.9
	.....	.....	.....	.....	6.50	1.68	18.55	21.43	.....	18.15	.....
	An_{46}36.86	An_{43}36.45	.....	An_{41}49.57	An_{46}64.80	An_{39}60.37	An_{28}53.52	An_{24}60.25	.....	An_{29}62.25	.....
	.....	.....	.....	.....	.....	0.25	.....	.....	.....	0.63	.....
	1.95	5.53	.....	1.39	0.75	17.60	16.20	13.32	.....	8.54	.....
	3.05	.....	.....	0.34	.....	.....	.....	.....	.....	.....	.....
	.....	.....	.....	0.20	.....	.....	0.09	0.07	.....	.....	.....
	2 73	.....	.....	11.03	13.67	.....	.....	2.75	.....	7.77	.....
	3.76	5.40	.....	6.26	8.02	11.27	4.98	1.50	.....	0.86	.....
	51.63	51.07	.....	28.71	3.96	7.40	2.67	.....	.....	.....	.....
	.....	.....	.....	.....	.....	.....	.....	.....	.....	1.18	.....
	.....	.....	.....	1.51	.....	.....	.....	.....	.....	.....	.....
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
	0.11	0.17	.....	0.64	1.79	0.91	.....	.....	.....	0.51	.....
	.....	.....	.....	il 0.17	.....	.....	.....	.....	.....	.....	.....
	.....	.....	.....	.....	mg. 0.21	.....	.....	.....	.....	.....	.....
	0.09	1.38	.....	0.18	0.29	0.52	0.35	0.68	.....	0.11	.....
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
	100.00	100.00	.....	100.00	.....	100.00	100.00	100.00	.....	100.00	.....
	.....	106.6	124.6	.....	137.0	142.0	226.2	257.6	261.0	.....	270.0
	.....	0.62	0.21	.....	0.02	0.12	0.14	0.11	0.21	.....	0.14
	.....	20.25	24.2	.....	31.80	30.00	34.4	36.75	37.18	.....	37.8
	.....	45.35	42.5	.....	31.10	32.35	27.2	24.65	25.28	.....	22.4
	.....	27.50	23.8	.....	30.21	22.05	17.2	14.42	15.72	.....	15.8
	.....	6.90	9.5	.....	6.89	15.60	21.2	24.18	21.82	.....	24.0
	.....	−21.0	−13.4	.....	99.44	−20.4	41.4	60.88	73.0	.....	64.0
	.....	0.11	0.14	.....	0.23	0.17	0.21	0.11	0.13	.....	0.20
	.....	0.60	0.55	.....	0.51	0.46	0.45	0.47	0.43	.....	0.42
	.....	0.61	0.56	.....	0.97	0.68	0.63	0.58	0.62	.....	0.71
	.....	.....	.....	.....	4.53	.....	13.15	17.21	20.59	.....	19.82
	.....	3.18	4.85	.....	0.56	8.98	11.59	6.59	6.98	.....	11.16
	.....	24.50	29.40	.....	22.90	41.80	41.20	48.11	41.90	.....	41.23
	.....	28.34	27.89	.....	44.90	25.30	17.12	15.02	18.00	.....	15.87
	.....	6.91	9.69	.....	10.22	.....	5.73	4.99	4.61	.....	3.84
	.....	4.21	7.44	.....	10.35	0.02	1.59	5.04	4.00	.....	3.77
	.....	9.73	4.43	.....	.....	6.64	.....	.....	.....	.....	.....
	.....	6.52	3.53	.....	.....	8.04	.....	.....	.....	.....	.....
	.....	10.80	6.87	.....	3.62	4.83	1.73	0.08	0.05	.....	0.73
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
	.....	2.78	1.93	.....	1.46	1.72	1.00	0.95	1.64	.....	1.34
	.....	0.72	2.11	.....	0.23	1.12	0.92	0.75	1.35	.....	1.14
	.....	1.38	0.36	.....	0.30	0.53	0.36	0.69	0.13	.....	0.16
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
	.....	0.93	1.40	.....	0.33	1.02	0.66	0.51	0.88	.....	0.92
	.....	100.00	99.90	.....	100.00	100.00	100.00	100.00	100.13	.....	99.98

6. Diorite, Merimasku
7. Hornblende charnockite, Askainen
8. Charnockite, Rymättylä
9, 9a. Charnockite, Merimasku
10, 10a. Charnockite, Merimas
11. Trondhjemite, Masku
12, 12a. Granite-trondhjemite,
13. Granite-trondhjemite, Livo

TABLE 1.—*Chemical and mineralogical composition of the rocks of the*
Including their Niggli numbers and norms.

	2a	3	4	4a	5	6	7	8	9	9a	
59	1059	1142	877	877	837	833	881	1087	834	834	
.72	50.05	48.92	51.12	52.40	53.37	53.25	63.59	66.40	66.31	65.21	
.86	0.37	0.38	1.11	0.31	0.12	0.59	0.51	0.39	0.71	0.27	
.06	16.41	15.73	16.81	16.71	21.00	19.05	16.36	16.05	16.02	16.66	
.86	1.83	1.91	1.33	1.44	1.00	1.18	0.69	0.66	1.13	0.68	
.07	8.70	7.80	8.00	8.30	6.11	6.63	4.28	3.34	3.29	4.08	
.18	0.32	0.28	0.13	0.18	0.08	0.12	0.07	0.05	0.04	0.03	
.22	7.22	8.35	6.44	6.58	4.12	3.75	2.31	2.01	1.86	2.55	
.64	10.62	11.69	9.12	9.45	10.95	7.73	4.48	3.45	3.72	3.88	
.63	3.01	2.88	3.46	3.72	2.69	4.93	4.84	5.67	4.94	5.39	
.51	0.49	0.54	0.82	0.19	0.09	1.52	1.95	1.11	1.18	0.83	
.19	0.04	0.59	0.16	0.08	0.13	0.23	0.16	0.30	0.04	0.05	
...	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
...	.....	.....	.....	0.08	.....	.....	.....	.....	.....	.....	.....
.18	0.77	0.74	1.20	0.50	0.16	0.74	0.57	0.43	0.65	0.28	
.30	0.17	0.19	0.20	0.14	0.10	0.28	0.19	0.14	0.23	0.09	
.42	100.00	100.00	99.90	100.00	100.00	100.00	100.00	100.00	100.12	100.00	9
...	.....	.....	.....	.....	6.50	1.68	18.55	21.43	.....	18.15	...
...	$An_{46}36.86$	$An_{42}36.45$	.....	$An_{41}49.57$	$An_{60}64.80$	$An_{30}60.37$	$An_{26}53.52$	$An_{24}60.25$	.....	$An_{28}62.25$	...
...	.....	.....	.....	.....	.....	0.25	.....	.....	.....	0.63	...
...	1.95	5.53	.....	1.39	0.75	17.60	16.20	13.32	.....	8.54	...
..	3.05	.....	.....	0.34	.....	.....	.....	.....	.....	.....	...
...	.....	.....	.....	0.20	.....	.....	0.09	0.07	.....	.....	...
...	2.73	.....	.....	11.03	13.67	.....	.....	2.75	.....	7.77	...
...	3.76	5.40	.....	6.26	8.02	11.27	4.98	1.50	.....	0.86	...
...	51.63	51.07	.....	28.71	3.96	7.40	2.67	.....	.....	.....	...
...	.....	.....	.....	.....	.....	.....	.....	.....	.....	1.18	...
...	.....	.....	.....	1.51	.....	.....	.....	.....	.....	.....	...
...	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	...
...	0.11	0.17	.....	0.64	1.79	0.91	.....	.....	.....	0.51	...
...	.....	.....	.....	il 0.17	.....	.....	.....	.....	.....	.....	...
...	.....	.....	.....	.....	mg. 0.21	.....	.....	.....	.....	.....	...
..	0.09	1.38	.....	0.18	0.29	0.52	0.35	0.68	.....	0.11	...
...	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	...
..	100.00	100.00	.....	100.00	.....	100.00	100.00	100.00	.....	100.00	...
7	.....	106.6	124.6	.....	137.0	142.0	226.2	257.6	261.0	.....	27
31	.....	0.62	0.21	.....	0.02	0.12	0.14	0.11	0.21	.....	
88	.....	20.25	24.2	.....	31.80	30.00	34.4	36.75	37.18	.....	3
41	.....	45.35	42.5	.....	31.10	32.35	27.2	24.65	25.28	.....	2
00	.....	27.50	23.8	.....	30.21	22.05	17.2	14.42	15.72	.....	1
71	.....	6.90	9.5	.....	6.89	15.60	21.2	24.18	21.82	.....	2
1	.....	−21.0	−13.4	.....	99.44	−20.4	41.4	60.88	73.0	.....	6
27	.....	0.11	0.14	.....	0.23	0.17	0.21	0.11	0.13	.....	
53	.....	0.60	0.55	.....	0.51	0.46	0.45	0.47	0.43	.....	
50	.....	0.61	0.56	.....	0.97	0.68	0.63	0.58	0.62	.....	
	.....	.....	.....	.....	4.53	.....	13.15	17.21	20.59	.....	1
93	.....	3.18	4.85	.....	0.56	8.98	11.59	6.59	6.98	.....	1
29	.....	24.50	29.40	.....	22.90	41.80	41.20	48.11	41.90	.....	4
90	.....	28.34	27.89	.....	44.90	25.30	17.12	15.02	18.00	.....	1
53	.....	6.91	9.69	.....	10.22	.....	5.73	4.99	4.61	.....	
35	.....	4.21	7.44	.....	10.35	0.02	1.59	5.04	4.00	.....	
90	.....	9.73	4.43	.....	.....	6.64	.....	.....	.....	.....	
16	.....	6.52	3.53	.....	.....	8.04	.....	.....	.....	.....	
0	.....	10.80	6.87	.....	3.62	4.83	1.73	0.08	0.05	.....	
3	.....	2.78	1.93	.....	1.46	1.72	1.00	0.95	1.64	.....	
2	.....	0.72	2.11	.....	0.23	1.12	0.92	0.75	1.35	.....	
3	.....	1.38	0.36	.....	0.30	0.53	0.36	0.69	0.13	.....	
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	
8	.....	0.93	1.40	.....	0.33	1.02	0.66	0.51	0.88	.....	
2	.....	100.00	99.90	.....	100.00	100.00	100.00	100.00	100.13	.....	9

6. Diorite, Merimasku
7. Hornblende charnockite, Askainen
8. Charnockite, Rymättylä
9, 9a. Charnockite, Merimasku
10, 10a. Charnockite, Me
11. Trondhjemite, Mask
12, 12a. Granite-trondhj
13. Granite-trondhjemite

he rocks of the charnockite and granite series
nbers and norms.

	9a	10	10a	11	12	12a	13	14	15	16	17
	834	835	835	1053	891	891	905	1153	1158	1161	1159
1	65.21	66.68	64.56	68.72	68.48	67.40	68.80	44.33	52.78	62.23	69.49
1	0.27	0.59	0.26	0.47	0.36	0.26	0.23	0.57	2.40	1.39	0.31
2	16.66	15.87	16.37	15.17	16.07	16.48	16.73	14.57	15.09	15.41	14.68
3	0.68	0.92	0.77	0.66	1.51	0.70	0.40	5.10	3.45	1.68	0.54
9	4.08	2.93	4.31	2.86	1.29	2.43	1.77	14.14	10.00	5.35	2.23
4	0.03	0.07	0.04	0.04	0.03	0.02	0.02	0.44	0.06	0.04	0.02
6	2.55	1.55	2.42	2.03	1.40	1.33	0.74	4.62	2.69	2.01	1.07
2	3.88	3.64	3.97	3.05	2.87	3.40	2.23	9.33	6.10	4.14	1.31
4	5.39	4.86	5.24	4.58	5.43	5.08	5.05	2.48	3.82	4.06	2.71
8	0.83	1.88	1.62	1.67	1.94	2.33	3.68	1.38	1.80	2.47	6.97
4	0.05	0.07	0.05	0.08	0.11	0.18	0.02	2.15	0.86	0.37	0.01
	.....	.....	.....	0.04	0.68	0.34	0.27	.....	0.09	0.17	0.26
	.....	.....	.....	.....	.....	.....	.....	.....	0.01	.....	.....
5	0.28	0.80	0.31	0.51	.....	.....	.....	0.75	0.67	0.53	0.31
3	0.09	0.12	0.10	0.12	0.12	0.05	0.06	0.18	0.18	0.15	0.09
2	100.00	99.98	100.02	100.00	100.29	100.00	100.00	100.00	100.01	100.00	100.00
	18.15	.....	16.60	29.60	.....	21.50	20.58	7.84	14.10	22.25	24.35
	$An_{28}62.25$	.....	$An_{28}58.30$	$An_{37}51.55$	.....	$An_{22}54.75$	$An_{26}52.70$	$An_{47}35.75$	$An_{32}45.80$	$An_{30}47.40$	$An_{21}28.21$
	0.63	.....	5.94	1.98	.....	10.43	17.82	.....	0.66	6.01	36.25
	8.54	.....	6.68	16.50	.....	6.15	7.75	16.53	20.83	17.82	10.39
	.....	.....	.....	.....	.....	.....	0.52	.....	.....	.....	.....
	.....	.....	.....	0.14	.....	.....	0.17	0.11	.....	.....	.....
	7.77	.....	5.02	.....	.....	.....	.....	.....	.....	.....	.....
	0.86	.....	3.08	.....	.....	.....	.....	.....	.....	.....	.....
	.....	.....	.....	.....	.....	5.92	.....	22.62	3.92	.....	.....
	1.18	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
	.....	.....	0.12	.....	.....	.....	.....	.....	.....	.....	.....
	0.51	.....	0.57	.....	.....	0.86	0.31	12.25	8.23	3.28	0.48
	.....	.....	.....	.....	.....	.....	.....	.....	4.37	2.13	.....
	.....	.....	.....	.....	.....	.....	.....	.....	0.03	.....	.....
	0.11	.....	0.09	0.17	.....	0.39	0.05	4.90	1.94	0.86	0.03
	.....	.....	.....	0.06	.....	.....	.....	.....	0.13	0.25	0.38
	100.00	.....	100.00	100.00	.....	100.00	100.00	100.00	100.00	100.00	100.00
	.....	270.0	.....	289.3	293.0	.....	314.0	98.70	149.0	223.0	330.0
	.....	0.14	.....	0.15	0.16	.....	0.08	0.95	5.10	3.76	0.11
	.....	37.8	.....	37.74	40.50	.....	45.2	19.10	25.10	32.61	41.05
	.....	22.4	.....	25.15	18.55	.....	10.6	51.15	42.50	31.58	18.55
	.....	15.8	.....	13.83	13.15	.....	11.0	22.40	18.50	15.96	6.7
	.....	24.0	.....	23.28	27.80	.....	33.2	7.35	13.90	19.85	33.7
	.....	64.0	.....	96.20	81.8	.....	81.2	−30.7	−6.6	43.6	95.2
	.....	0.20	.....	0.19	0.11	.....	0.32	0.27	0.23	0.29	0.61
	.....	0.42	.....	0.51	0.48	.....	0.48	0.30	0.27	0.34	0.41
	.....	0.71	.....	0.55	0.71	.....	1.04	0.44	0.43	0.50	0.36
	.....	19.82	.....	24.17	21.12	.....	18.10	.....	4.85	15.60	20.96
	.....	11.16	.....	9.87	11.50	.....	21.80	8.21	10.74	14.61	41.30
	.....	41.23	.....	38.83	46.08	.....	42.92	21.02	32.45	34.42	23.00
	.....	15.87	.....	15.19	13.56	.....	11.02	24.42	18.45	16.50	6.46
	.....	3.84	.....	5.04	3.47	.....	1.84	5.29	6.67	4.99	2.65
	.....	3.77	.....	4.00	0.57	.....	2.57	9.97	11.66	6.23	3.16
	.....	.....	.....	.....	.....	.....	.....	4.35	.....	.....	.....
	.....	.....	.....	.....	.....	.....	.....	9.03	.....	.....	.....
	.....	0.73	.....	.....	.....	.....	.....	3.33	2.63	0.70	.....
	.....	.....	.....	0.24	0.05	.....	0.37	.....	.....	.....	0.29
	.....	1.34	.....	0.95	2.20	.....	0.58	7.40	5.00	2.43	0.79
	.....	1.14	.....	0.89	0.68	.....	0.44	1.08	4.56	2.66	0.59
	.....	0.16	.....	0.19	0.26	.....	0.03	4.97	2.01	0.87	0.03
	.....	.....	.....	.....	.....	.....	.....	.....	0.13	0.31	0.38
	.....	0.92	.....	0.63	0.80	.....	0.33	0.93	0.85	0.68	0.40
	.....	99.98	.....	100.00	100.29	.....	100.00	100.00	100.00	100.00	100.00

Charnockite, Merimasku
dhjemite, Masku
Granite-trondhjemite, Merimasku
ite-trondhjemite, Livonsaari
14. Gabbro, Räntämäki
15. Diorite, Räntämäki
16. Granodiorite, Räntämäki
17. Granite, Räntämäki

ju
se
se

li
T
re
id
in
cl
K
re
ra
th

ai
(A
sc
or
sou
por
2 k

T
opt
tha
thei

C
bler

H
equi
occu

T
biot
dark
± 0.
= 1
γ =
horn
occu

Th
posit
are la
miner
analy
area.

just as potash feldspar commonly does (Hietanen, 1943, p. 96). This sodium metasomatism caused by trondhjemitic solutions is comparable to that described by Goldschmidt (1920) from the Stavanger area in Norway.

Trondhjemites associated with the above rocks are even- and medium-grained, light bluish gray, and closely resemble the normal trondhjemites of the Kalanti area. The trondhjemite near the contact zone of the Vehmaa rapakivi (loc. 853) is probably related to this particular differentiation series. The plagioclase (An_{25}) in it is partly idiomorphic and more calcic than in the normal trondhjemites. Dark fine-grained inclusions are probably fragments of the older rocks and consist of rounded plagioclase grains surrounded by abundant small biotite flakes. The trondhjemites of Kuntikari, the analysis of which was given earlier (Hietanen, 1943, p. 44), greatly resembles this trondhjemite. The island Kuntikari is on the west side of the Vehmaa rapakivi pluton. The trondhjemitic rocks may have been widely distributed also in the areas now occupied by rapakivi.

The structure of a small diorite body about 1 km. north of Nummenkylä, Nousiainen, is comparable to that of the rocks in Livonsaari. Part of the plagioclase (An_{38}) occurs in larger partly rounded grains, and the rock looks porphyritic. Microscopic examination shows that these plagioclase "phenocrysts" consist of one large or several smaller grains. Biotite is the only dark mineral in this rock which on the south side of the lens grades over to a light-bluish-gray trondhjemite with the same porphyritic structure. A similar porphyritic diorite was also seen in a road cut about 2 km. north of Karjala church.

The composition of the hornblendites, gabbros, diorites, and trondhjemites, and the optical properties of their minerals, suggest that their chemical composition is near that of the corresponding rocks of the Kalanti area. Therefore the discussion of their petrography is not repeated here.

Charnockite series.—The basic division of the charnockite series includes hornblendites, gabbros, diorites, and norites.

Hornblendites and gabbros are megascopically and mineralogically similar to the equivalent rocks of the trondhjemite series. The only difference appears in the local occurrence of some diopside (*e.g.*, in no. 841, Pl. 1).

The minerals in the hornblendite no. 1140 (Pl. 1) are hornblende, plagioclase (An_{49}), biotite, apatite, magnetite, and ilmenite. The hornblende shows pleochroism γ = dark brownish green, β = grayish green, α = pale green; refractive indices $\gamma = 1.682 \pm 0.001$, $\alpha = 1.661 \pm 0.001$, and $\gamma \wedge c = 21°$; these properties indicate [FeO]:[MgO] = 1 according to Winchell. Biotite has $\gamma = \beta = 1.650 \pm 0.001$; the pleochroism is $\gamma = \beta$ = dark reddish brown, α = pale tan. Biotite grows in long laths through the hornblende and plagioclase grains and also cuts their contacts. Ilmenite inclusions occur in the hornblende.

The rock was analyzed by means of the integration stage, and the chemical composition calculated. The chemical compositions of the minerals used for this purpose are largely based on the mineral analysis of the rocks of the Kalanti area. A new mineral analysis of hypersthene was made by Mr. O. von Knorring (Table 2). The analysis of hornblende was calculated from the rock analysis no. 18 in the Kalanti area. The ratio [FeO]:[MgO] was determined in each case by means of the optical

properties of the minerals and Winchell's tables. The geometric analysis and the calculated chemical composition of the hornblendite 1140 are shown in Table 1 (no. 1). Gabbro no. 1142 was analyzed similarly (Table 1, no. 3). It contains in addition to the minerals of the hornblendite 1140 also some diopside and quartz. The plagioclase is An_{42-43}.

TABLE 2.—*Chemical analysis of hypersthene no. 1059*

	%	Mol.
SiO_2	48.94	8116
TiO_2	trace	
Al_2O_3	1.66	162
Fe_2O_3	5.28	331
FeO	23.60	3285
MnO	0.42	59
MgO	16.96	4206
CaO	1.11	198
H_2O	0.35	
	98.32	

There are two types of rocks which could be called norites: (1) true norites, directly related to the intermediate charnockites and representing, with the gabbroic rocks, the basic division of the charnockite series; and (2) the small lenticular bodies which occur in the kinzigites parallel to the bedding.

The true norites are medium- to coarse-grained and vary from brownish black to brown and gray. The common brownish tint, caused by the presence of abundant hypersthene, permits ready recognition in the field.

The basic norites are megascopically very similar to the gabbros. They are dark and coarse- to medium-grained. Hornblende is prominent among the dark constituents. Rhombic as well as monoclinic pyroxenes, and biotite occur only in minor amounts. In some localities, however, coarse-grained hypersthene has accumulated into 1 to 5 m long and 20 to 80 cm broad masses. Such a hypersthene from a small norite lens about 4 km south of Masku, locality no. 1059, was separated by means of Clerici solution. Chemical analysis was made by Mr. O. von Knorring (Table 2).

The ratio is fs:en = 42:58 according to the analysis. The indices of refraction measured by the immersion method are $\gamma = 1.733 \pm 0.001$, $\beta = 1.730 \pm 0.001$, $\alpha = 1.719 \pm 0.001$. The comparison of these values with the chemical composition shows a relation between the chemistry and optics similar to that indicated in Winchell's table—[FeO]:[MgO] = 41:59. The refractive indices of the hypersthene in the main part of the same rock are $\gamma = 1.728 \pm 0.001$, $\beta = 1.725 \pm 0.001$, $\alpha = 1.713 \pm 0.001$; and $-2V = 48°$.

Diopside shows $\gamma = 1.715 \pm 0.001$, $\beta = 1.696 \pm 0.001$, $\alpha = 1.690 \pm 0.001$, which indicates according to Winchell a composition of 33 per cent diopside and 67 per cent hedenbergite. Hornblende has the common pleochroism: γ = bluish green, β = green, α = light yellowish green. The refractive indices are $\gamma = 1.691 \pm 0.001$, $\beta = 1.682 \pm 0.001$, $\alpha = 1.673 \pm 0.001$. $\gamma \wedge c = 21°$. These properties indicate a ratio [FeO]:[MgO] = 2:1.

The biotite occurs in narrow laths, which penetrate the other minerals and transect their contacts. The pleochroism is γ = brown, α = light yellowish brown, and the index of refraction $\gamma = 1.647 \pm 0.001$. Table 1 shows the chemical analysis (2) and the chemical composition calculated on the basis of the geometric analysis (2a) of this norite.

TABLE 3.—*Calculated chemical composition of hornblende*

	Per cent		Per cent
SiO_2	45.2	MgO	14.0
TiO_2	0.4	CaO	13.4
Al_2O_3	10.0	Na_2O	0.5
Fe_2O_3	3.0	K_2O	0.1
FeO	11.7	H_2O+	1.0
MnO	0.5	H_2O-	0.2
			100.0

The norite 877 which occurs as a border facies of the charnockites is coarse- to medium-grained and dark (Pl. 3, fig. 1). A part of the hypersthene has crystallized as larger grains or gathered into small masses. The refractive indices are $\gamma = 1.734 \pm 0.001$, $\beta = 1.730 \pm 0.001$, $\alpha = 1.720 \pm 0.001$. The diopside has $\gamma = 1.718 \pm 0.001$, $\beta = 1.704 \pm 0.001$, $\alpha = 1.691 \pm 0.001$. The hornblende with $\gamma = 1.690 \pm 0.001$ and $\gamma \wedge c = 16°$ is the dominant dark constituent. Biotite with $\gamma = 1.660 \pm 0.001$ transects the other minerals as long narrow laths. Some actinolite with $\gamma = 1.665 \pm 0.001$ occurs as an alteration product of hypersthene.

For comparison of the geometrical and chemical methods specimen 877 was also measured with the integration stage. Mr. O. von Knorring analyzed it chemically. Both analyses are shown in Table 1 (no. 4, 4a) and are practically the same. The sum of the alkalies is about the same. The larger amount of Na_2O in the calculated analysis is obviously because plagioclase contains a certain amount of K_2O. This result proves that geometric analyses can be successfully used instead of chemical analyses. The method involves a determination of the true composition of the minerals, either chemically or on the basis of their optical properties. In this particular case the analyses of biotite and diopside from the Kalanti area and the analysis of hypersthene no. 1059 were used, after their ratio of [FeO]:[MgO] was changed to correspond to the indices of refraction of the minerals in no. 877. The composition of hornblende calculated from the analysis of no. 18 in Kalanti area was used (Table 3).

The ratio [FeO]:[MgO] calculated from the analysis is about 1:2. The optical properties correspond approximately with the same ratio according to Winchell's tables.

The small lenticular bodies of norite are fine-grained and dark or gray. The minerals are quartz, plagioclase, hypersthene, diopside, hornblende, biotite, and the accessories apatite, magnetite, and zircon. Much of the hypersthene forms huge holoblasts, which include numerous partly rounded quartz and feldspar grains. The hornblende and pyroxenes occur in separate layers. Variation in the mineral com-

position is largely due to the absence of diopside or hornblende and to variations in the composition of the plagioclase. In the diopside- and hornblende-bearing varieties the plagioclase is more calcic (*e.g.*, in nos. 837, 1306, 1315 it is An_{65-70}) than in the hypersthene-biotite norites (*e.g.*, in nos. 1145, 1152, 1258, An_{46-55}). Chemically this indicates more lime, both salic and femic, in the diopside- and hornblende-bearing varieties, which also contain less quartz and are thus more basic than the hypersthene-biotite norites. A small lenticle north of the Merijärvi charnockite pluton (no. 837) represents this type. It contains abundant rhombic and monoclinic pyroxenes and only very little quartz as small rounded grains. The other minerals are plagioclase (An_{65-70}), hornblende, biotite, apatite, magnetite, pyrrhotite, and zircon.

The hypersthene has a higher birefringence than in the charnockites: $\gamma - \alpha = 0.016$. The indices of refraction are $\gamma = 1.728 \pm 0.001$, $\beta = 1.724 \pm 0.001$, $\alpha = 1.712 \pm 0.001$, and the optic angle $-2V = 43°$. Pleochroism is lacking. A few grains show an alteration rim of cummingtonite. Diopside appears very similar to the hypersthene. It can be distinguished only by its extinction angle ($\gamma \wedge c = 42°$). The optic angle is $+2V = 50°$. The indices of refraction are $\gamma = 1.715 \pm 0.001$, $\beta = 1.697 \pm 0.001$, $\alpha = 1.687 \pm 0.001$. The hornblende is of a common brown type and has $\gamma = 1.676 \pm 0.001$, $\beta = 1.670 \pm 0.001$, $\alpha = 1.656 \pm 0.001$. Biotite has $\gamma = \beta = 1.634 \pm 0.001$.

The mineral analysis and the calculated chemical composition of this rock is shown in Table 1 (no. 5).

The si value is lower than in the normal gabbros and higher than in the normal diorites; fm is smaller than in normal norites, al and c larger. This is due to the presence of abundant calcic plagioclase. The composition of this rock is near the anorthositic norites described by Buddington (1939) in the Adirondacks area. There is 28.4 per cent mafic minerals in the present rock.

The norites 1306 and 1315 are very similar to the norite no. 837. Also they contain abundant calcic feldspar and diopside or hornblende in addition to hypersthene and are thus rich in lime. Chemically they are near the pyroxene amphibolites. Their structure and the mode of occurrence also are very similar to the sedimentary hypersthene-bearing rocks. On the other hand there is no fundamental difference between them and the more acidic norites, which are clearly associated with the charnockites. No. 1152 is an example of this type. It occurs as a small lenticle in the kinzigite. At weathered surfaces it appears very similar to the charnockites but is finer-grained. The hypersthene occurs in large holoblasts, which include rounded quartz and feldspar grains. The hornblende also tends to form larger grains which include small quartz grains. The plagioclase has a composition of An_{49-53}.

The long, narrow occurrences east of Raisio church and 4 km. south of Turku appear more like beds. They are fine-grained, dark gray, and are intercalated with the amphibolites and biotite leptites rich in calcic plagioclase. The hypersthene-bearing layers are 1 to 3 meters thick. Microscopic examination shows that also in these layers hypersthene occurs in large holoblasts, and, if hornblende is present, it occurs in separate narrow layers.

The fine-grained norite no. 875, which occurs as a vein in the coarse-grained diorite on the southern contact zone of the Merijärvi charnockite pluton and is thus con-

sidered magmatic, is mineralogically and structurally close to the norites nos. 837 and 1152. In addition no. 875 shows granoblastic structure, and hypersthene tends to form larger grains. Thus it is difficult to decide whether the norites just described should be considered magmatic or sedimentary. Their occurrence in long narrow lenses parallel to the bedding may result either from bedding or from veining. Possibly both types occur. The chemical composition of those rich in calcic minerals, and also layering as described in the norites east of Raisio and 4 km south of Turku, would suggest a sedimentary origin. In their mineral assemblage and often in structure they resemble those norites which grade into charnockites. It is therefore impossible to separate the two groups. The interesting fact is that the same mineral assemblage may be derived either by metamorphic processes or by crystallization from magma.

Diorites occur as a border facies of the charnockite plutons and as small lenses in the kinzigites. They are closely associated with the gabbros. Most diorites are exceptionally dark due to the abundance of mafic minerals and the bluish color of the salic constituents. The main constituents are plagioclase (An_{36-37}), quartz, and biotite. Hornblende and pyroxenes may occur.

Specimen 904 is an example of the dark biotite diorite, which grades into the foliated border facies of diorite-trondhjemite. The structure in these diorites is granoblastic, as generally in all intrusive rocks of this area.

Diorite no. 1143, from the same locality as hornblendite no. 1140 and gabbro 1142, contains biotite aggregates, which are clearly an alteration product after hypersthene. Some small diopside grains are associated with these aggregates. Plagioclase is An_{40}, as is common in the diorites. The amount of hornblende approximately equals that of biotite.

The diorite grades into a hypersthene-bearing charnockite and completes the differentiation series hornblendite → gabbro → diorite → charnockite.

The rocks of intermediate division mapped as charnockites are medium-grained, gray or brownish. More than 50 per cent of the rock is plagioclase (An_{25-33}). Quartz is about 20 per cent. Microcline may be present. The dark minerals are hypersthene, diopside, hornblende, and biotite. Rarely do all occur in the same specimen, but there are usually two or three mafic minerals present in the rock. This gives a great variation to the intermediate division. We can talk about normal charnockites, trondhjemitic charnockites, and hornblende charnockites. The texture is always granoblastic. Normal charnockites contain microcline in addition to plagioclase and quartz. It occurs as antiperthitic inclusions in plagioclase and in small scattered grains. Rhombic and monoclinic pyroxenes, and biotite are the dark constituents. Apatite and magnetite are the common accessories (Table 1, analyses 10, 10a). The percentage of microcline is about the same as that of the biotite and hypersthene, while the amount of plagioclase is 10 times greater. The chemical analysis shows $2\frac{1}{2}$ times more N_2O than K_2O.

Trondhjemitic charnockites contain only very little or no microcline. Their composition thus approaches the intermediate members of the trondhjemite series, and they grade into trondhjemites. Normal charnockites also grade into trondhjemitic types, as seen in the charnockite intrusion south of Lake Kuusniemenjärvi, Meri-

masku. The center of this pluton consists of the brownish, microcline-bearing charnockite (spec. no. 835, Anal. 10 in Table 1) and grades into a light-gray rock which contains microcline only as antiperthitic inclusions (spec. no. 834, Anal. 9 in Table 1; Figs. 9, 10). Toward the margins the plagioclase becomes more acidic, the pyroxenes disappear, and the rock changes into a light-bluish-gray trondhjemite. The minerals of the trondhjemitic charnockites are the same as those of the normal charnockites

TABLE 4.—*Relation of SiO_2 content to refractive indices of hypersthene and diopside*

In per cent

Spec. number	1059	877	1058	834
SiO_2	47.72	51.12	ca. 55.0	66.31
γ hypersthene	1.728	1.734	1.736	1.738
γ diopside	1.715	1.718		1.720

except for microcline. The plagioclase, determined by Reinhard's (1931) method, varies from An_{25} to An_{28}. The hypersthene shows no pleochroism, its optic angle is $-2V = 54.5°$, and the indices of refraction are $\gamma = 1.738 \pm 0.001$, $\beta = 1.734 \pm 0.001$, $\alpha = 1.724 \pm 0.001$, indicating hypersthene of en_{58} fs_{42} according to Winchell. The alterations of this mineral are described separately. The diopside grains are smaller and fewer than the hypersthene. It appears clear and unaltered under the microscope. The optical properties are: $+2V = 55°$; $\gamma = 1.720 \pm 0.001$; $\beta = 1.709 \pm 0.001$; $\alpha = 1.702 \pm 0.001$, indicating a diopside with 40 per cent hedenbergite. The biotite is strongly pleochroic with $\gamma = \beta$ brown and α pale yellowish; the refractive index $\gamma = 1.650 \pm 0.001$. Symplectitic intergrowths with quartz are common. Figure 12 shows biotite myrmekite along the margins of a hypersthene, which is partly altered to biotite.

The refractive indices of pyroxenes seem to rise regularly with increasing silica content in the rock. Specimens 1059, 877, 1058, and 834 form a series from basic norites to charnockites. Table 4 shows their weight percentage of silica and the refractive index γ of the hypersthene and diopside.

According to Winchell this suggests an increase of the iron content in pyroxenes toward the acidic members of this rock series.

The margins of the long narrow charnockite lens south of Masku consist of a dark-gray charnockite in which the dark constituents are hypersthene, diopside, and biotite. Hornblende is found occasionally. The amount of biotite varies. As a rule the central part is more acidic. It is light brownish gray resembling charnockite no. 834 (Anal. 9 in Table 1).

The basic norite no. 1059 and the norite no. 1058 occur as long lenses in this charnockite. The contacts between the two rocks are gradational.

Hypersthene is rarely fresh like diopside but usually shows alterations into various minerals. A serpentinelike green mineral along the cracks of hypersthenes is seen in all the rocks of this series, but it is far more common in the acidic division. In the dioritic members the cracks are narrow and very similar to those in the diopside in the same rock. In the granodioritic members the cracks of the hypersthene are broad and filled with various alteration products, while the diopside is fresh (Fig. 7).

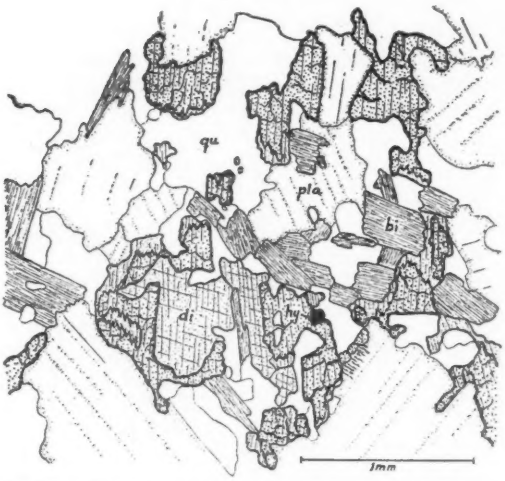


FIGURE 7.—*Unaltered diopside (di) surrounded by hypersthene (hy), which changes into cummingtonite* Charnockite south of the Lake Kuusniemenjärvi, Rymättylä (no. 834, Pl. 1).

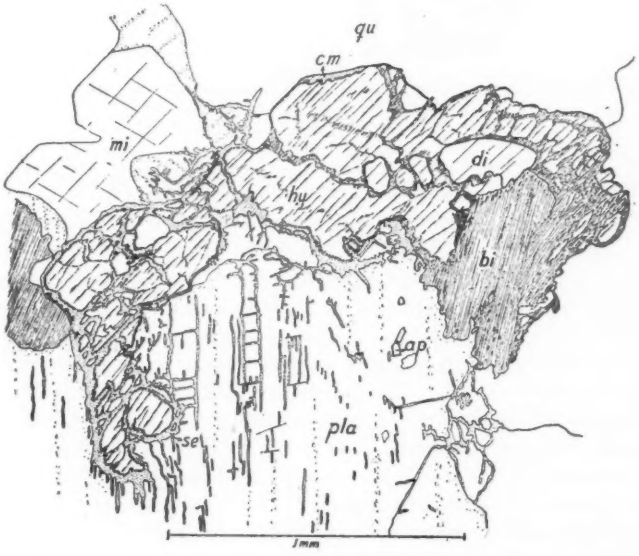


FIGURE 8.—*Alteration of hypersthene (hy) to biotite (bi) along the margins and to serpentine (se) along the cracks*

Serpentine (antigorite) occurs also in the cleavage cracks of the neighboring plagioclase (pla). Charnockite ½ km south of Lake Kuusniemenjärvi, Rymättylä (no. 835).

Figure 8 shows such alteration on a larger scale. On the left side are remnants of hypersthene surrounded by a pale-green mineral, which also fills the cleavage cracks

of the adjacent plagioclase and occurs as marginal films between quartz and feldspar grains. This mineral is composed of fine fibers, which are characterized by weak pleochroism: γ = green, α = pale green, positive elongation, extinction angle of 0° and birefringence of about 0.01. The mineral is most likely a variety of antigorite.

FIGURE 9.—*Alteration of hypersthene (hy) to cummingtonite (cm) in its margins*
Charnockite south of Lake Kuusniemenjärvi (834).

A portion of the iron oxides in the hypersthene now forms small magnetite grains among the antigorite. This alteration may be expressed as follows:

$$3\,(FeO\cdot 2MgO\cdot 3SiO_2) + 4H_2O + O \rightarrow 2(2H_2O\cdot 3MgO\cdot 2SiO_2) + 5SiO_2 + FeO\cdot Fe_2O_3$$
3 hypersthene + 4 water → 2 antigorite + 5 quartz + magnetite

The change from hypersthene into serpentine has in no case advanced further than this. Where hypersthene has been destroyed the end products are cummingtonite or biotite.

1. Alteration to cummingtonite seems to take place in rocks poor in potash feldspar. (*See* Figures 9 and 10.) Microcline occurs only in small antiperthitic inclusions in the plagioclase. Figure 9 shows hypersthene changing into cummingtohite in its margins. The center contains cracks which are filled with a green mineral, probably antigorite. The narrow zone between this central part and the cummingtonite margin is also greenish and has a lower index of refraction than the hypersthene and the cummingtonite. Its extinction is like that in hypersthene. It is therefore thought to be somewhat altered hypersthene. The marginal zone is pale-greenish cummingtonite, whose cleavage cracks are parallel to those in the hypersthene. The angle of extinction of this cummingtonite is 20°; +2V = about 70°; pleochroism is weak: γ = pale green, $\beta = \alpha$ = very pale yellowish green. The reaction of the alteration may be expressed as follows:

$$7(Fe, Mg)SiO_3 + SiO_2 + H_2O \rightarrow H_2O \cdot (Fe, Mg)_7Si_8O_{24}$$
7 hypersthene + quartz + water → cummingtonite

The margin of a hypersthene crystal in Figure 10 consists partly of cummingtonite and partly of diopside. The diopside is probably primary, formed as a reaction product between the hypersthene and the residual magma containing CaO, and not an

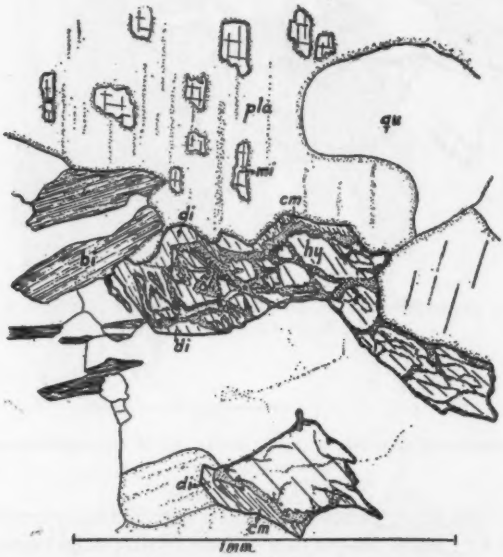


FIGURE 10.—*Hypersthene (hy) surrounded by primary diopside (di) and secondary cummingtonite (cm)* Location as in Figure 9.

alteration product like the cummingtonite. In the smaller grain below, a similar diopside rim can be seen on the left. The hypersthene is altered into cummingtonite along its margins, and there is a cummingtonite zone also between hypersthene and diopside. In the upper hypersthene grain alteration of that part which is surrounded by diopside is advanced only to the greenish stage. The diopside in both cases is unaltered.

Alteration of hypersthene into cummingtonite has been observed by several authors. Asklund (1925) has described it in a noritic gabbro in Stavsjö at Kolmården, and he lists a number of other authors. Lundegårdh (1943, p. 340) has found similar alteration rims of colorless fibrous amphibole around hypersthene remnants in the ultra-basic norites (hypersthene eucrites) in the Grovstanäs region, Sweden, and concurs with Asklund's identification of cummingtonite.

2. In the rocks containing abundant microcline the hypersthene has changed directly into biotite. Figure 11 shows various stages of this alteration. The main parts of the grains have broad brown margins, and the cleavage cracks are brown. In the narrow zone between the large biotite and quartz are remnants of hypersthene

surrounded by the same brown mineral which in places shows the optical properties of biotite. This biotite is somewhat lighter in color than the primary biotite, and the flakes are small. In Figure 8 a similar alteration of hypersthene into biotite can be seen in the right of the grain. In addition the cracks are filled with small biotite

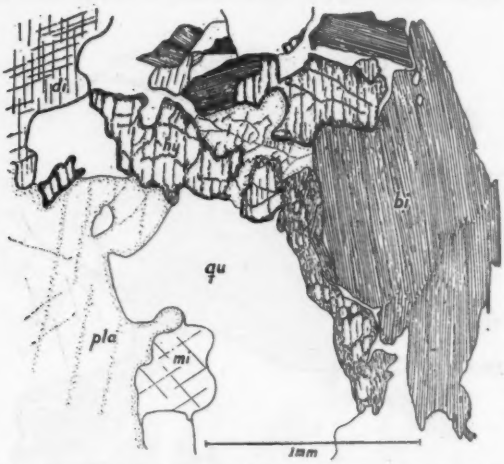


FIGURE 11.—*Alteration of hypersthene (hy) to biotite (bi) in microcline-bearing charnockites*
Location as in Figure 8.

flakes with their elongation parallel to the cleavage of the hypersthene. (*See also* Figure 2 of Plate 3.) In the central part of some broad cracks there occurs greenish serpentine similar to that which fills the cracks in the left of the same hypersthene grain. These cracks filled with green serpentine can still be seen when the hypersthene has completely altered into biotite—*e.g.*, in the charnockite south of Naantali and no. 1087 in Rymättylä (Fig. 13; Pl. 3, fig. 3). The biotitization seems therefore to be later than the alteration into serpentine. It may well be connected with the metasomatism of potash feldspar, in connection with regional migmatization, which is common in this area. Metasomatic alteration of hypersthene can be expressed as follows:

$$6(Fe, Mg)SiO_3 + 2KAlSi_3O_8 + 2H_2O \rightarrow K_2(Fe, Mg)_6(OH)_4(Si_6Al_2O_{20}) + 6SiO_2$$
6 hypersthene + 2 microcline + 2 water → biotite + 6 quartz

Abundant quartz is set free according to this equation. This quartz in many cases forms biotite-myrmekite with the secondary biotite. Figure 12 shows a border zone of a hypersthene grain with this sort of reaction rim. The biotite in the lower part of the figure is a common strongly pleochroic brown type with a distinct cleavage. The margins and the cracks of the hypersthene consist of the same brown mineral, but the narrow zone between the margin and the main part of the biotite—the zone of finer myrmekite—is lighter brown or greenish.

In the charnockite about 2 km east of Masku church the hypersthene shows all

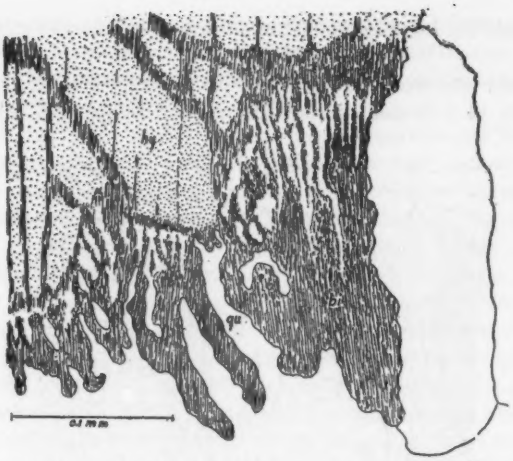


FIGURE 12.—*Biotite (bi) myrmekite around a hypersthene (hy) grain*
Location as in Figure 8.

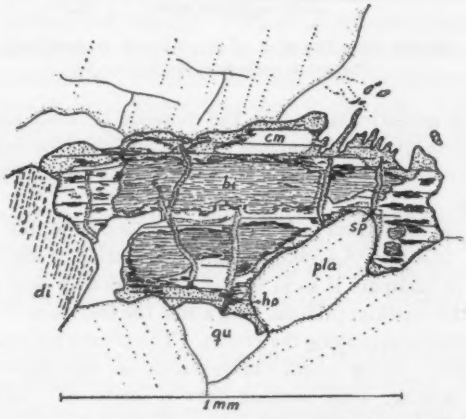


FIGURE 13.—*Hypersthene altered to serpentine (sp) and cummingtonite (cm) along the margins and to biotite (bi) in the center*
Charnockite 2 km east of Masku (no. 1032).

these alteration products. The outer zones have altered to cummingtonite and hornblende. The hornblende occurs regularly along the margins and was obviously formed as a reaction product between hypersthene and anorthite in a manner similar to that assumed by Eskola (1914) and Asklund (1925, p. 23). After that the inner zone altered to cummingtonite, and at last the center into biotite (Fig. 13). Figure 4 of Plate 3 shows a similar alteration, with unaltered hypersthene still in the center. In Figure 5 of Plate 3 the relations between the common green hornblende and hypersthene are not clear. Hypersthene, altering into serpentine and biotite along the

cracks, surrounds the hornblende. There is a zone of colorless amphibole between them.

Diopside alters into green hornblende in the pyroxene-bearing diorite-trondhjemites. In Figure 14 a diopside grain is surrounded by green hornblende, and the

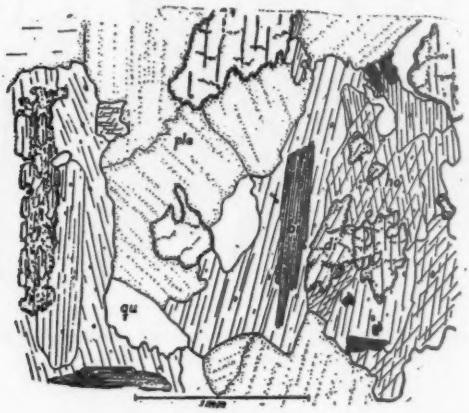


FIGURE 14.—*Alteration of diopside (di) to hornblende (ho)*
Charnockite southwest of Merijärvi (no. 881).

transition zone is gradational so that the diopside becomes light greenish. With advancing alteration the color becomes deeper, and the cleavage of hornblende appears. Almost every diopside grain in these rocks has greenish spots showing beginning of alteration.

Chloritization of the biotite is found in every rock type, but only locally and in small amounts. It is always accompanied by separation of small rutile needles.

Sericitization of feldspars is local and occurs especially near the contacts of the pegmatites and the rapakivi plutons. As a rule the feldspars are well preserved and contain only a little sericite along the cracks. The potash feldspar appears especially clear and unaltered. In the trondhjemite-granite no. 905, however, a part of it contains abundant sericite, while another part is unaltered. The unaltered part is thought to be secondary.

All the alterations are due to hydrothermal solutions at low temperatures and pressures.

Hornblende charnockites contain abundant hornblende and biotite, whereas hypersthene is lacking or has altered to biotite. Diopside occurs in minor amounts. The quality and quantity of the femic minerals present in the rock testify to the consanguinity among the various types. A section through the charnockite pluton around Lake Merijärvi affords a good example of this variation. The main part of this pluton consists of the brownish-gray, medium-grained rock, in which the amount of the femic minerals increases toward the margins. The central part consists of a more acidic type in which the biotite is the dominant and diopside the other dark mineral. Further, in the core of this pluton east of Lake Merijärvi is a light-brownish-gray trondhjemitic rock.

The border facies of the pluton is a medium-grained, well-foliated dark-gray rock, very similar to the hornblende trondhjemites of the Kalanti area. Microcline is usually lacking. Hornblende and biotite are abundant, diopside occurs in minor amounts. Plagioclase varies from An_{22} to An_{25}.

The brownish-gray main species contains less hornblende and biotite than the darker border facies, and most of the minerals show random orientation. Diopside may be present. In general the plagioclase in the diopside-bearing varieties shows a higher content of anorthite (An_{25-28}). Microcline is scarce and occurs as a film between the other minerals and as antiperthitic inclusions in the plagioclase in a manner similar to the trondhjemites of Kalanti. The hornblende is strongly pleocroic: γ = blue-green, β = green, α = light greenish yellow. The amount of biotite is perhaps a little larger than that of the hornblende. It is often chloritized and contains small grains of ore along the cleavages and margins. The geometric analysis of this type is shown in Table 1 (no. 7). The core of this pluton consists of a light-brownish-gray rock in which the main minerals are plagioclase (An_{22}), quartz, and biotite. The amount of microcline is about 10 per cent, larger than in the other rock types of the same pluton.

The series diorite → charnockite → hornblende trondhjemite → trondhjemite is well represented in the intrusives east of Masku. Diorites are coarse-grained, dark gray, and contain abundant hornblende. They grade into hornblende trondhjemites or charnockites. On the other hand the charnockites grade into hornblende trondhjemites. The charnockites are brownish as usual, and the bluish-gray hornblende trondhjemites can be well distinguished from them in the field. The minerals in the hornblende trondhjemites are quartz, plagioclase (An_{22-30}), hornblende, biotite, microcline, sphene, apatite, and magnetite. The amount of microcline varies greatly, and at places a part of it seems to be secondary.

The acidic rocks associated with the charnockites are either trondhjemitic, granite-trondhjemitic, or granitic, and they may contain, in addition to biotite, diopside, hypersthene, or garnet as a dark constituent. The acidic rocks, in which the amount of potash feldspar ranges from 20 to 30 per cent of the total feldspar, are called granite-trondhjemites.

In the Merimasku-Luonnonmaa pluton the bulk of the rock is similar to the intermediate zone of the Merijärvi pluton. In two zones, however, the rock is finer-grained and lighter-colored. Microscopic examination shows that it contains more microcline than any of the types mentioned above. The femic minerals, hornblende and biotite, occur in smaller quantities. Plagioclase (An_{28}), measured by the universal stage according to Reinhard's method, is more acidic than in the main part of the pluton.

The rock was analyzed chemically by Mr. O. von Knorring and also measured by the integration stage. The results are shown in Table 1 (no. 12 and 12a). Na_2O and K_2O percentages are very similar to those in the trondhjemites. The integration-stage analysis shows 10 per cent microcline, and accordingly the amount of K_2O is larger in the calculated analyses than in the chemical one. This is obviously due to the inhomogeneity of the rock.

The structure and the color of the granite-trondhjemite in Livonsaari, Askainen (no. 905; Pl. 3, fig. 6), resembles that of the fine-grained trondhjemite of Haidus

(Hietanen, 1943, p. 49). A minor portion of the quartz, however, appears as larger grains, like the quartz in the dark-gray trondhjemite east of Sundholm in the Kalanti area. The biotite with $\gamma = 1.660 \pm 0.001$ is the only dark constituent, and the plagioclase is An_{20}. The calculated analysis of this rock is shown in column 13 of Table 1.

The si value approaches the Haidus trondhjemite; al and alk values are somewhat larger, and fm smaller. The k value shows the essential difference between these two rocks: it is 0.32 in no. 905. The same value in the normal trondhjemites is about 0.20 and in the granites 0.42.

The narrow granite-trondhjemite body north of the Merimasku charnockite pluton consists of fine-grained light-gray rock with a brownish tint. Biotite is scarce and not well oriented. The potash content is approximately the same as in no. 905. The light-bluish-gray granite-trondhjemite of Tammisaari (no. 884) has a more trondhjemitic composition. The content of potash feldspar is somewhat smaller. The composition of the plagioclase in no. 888 and 884 is An_{22-23}.

The hornblende trondhjemites of Masku grade into hornblende-bearing granite-trondhjemites and trondhjemites. The former are bluish gray and medium-grained rocks, in which the amount of microcline varies from 20 to 35 per cent of the total feldspar. The plagioclase is An_{22-25}. Hornblende and biotite are the dark constituents, and sphene, apatite, and zircon the accessories.

Toward the northeast, the hornblende disappears gradually, and the rock becomes lighter and grades into a trondhjemite in which the amount of the microcline ranges from 5 to 20 per cent. The main part of this trondhjemite is granoblastic and shows good foliation. However, at some localities—*e.g.*, no. 1067—part of the feldspar occurs as larger, partly idiomorphic thick laths. Toward the east these rocks grade into the diorite of Pöytyä.

The reddish-gray granite-trondhjemites of the northern part of the area are mineralogically similar to those described above, but their genesis is different. They are partly granitization products of hornblende-bearing volcanic rocks as described under "Granitization" and partly they represent the acidic members of the Pöytyä diorite.

The long narrow satellitic intrusions in the kinzigites near the contact zones of the charnockite plutons consist exclusively of trondhjemites. Their color varies. Those in Luonnonmaa and in the southern Merimasku are light bluish gray, while the smaller lenses of Masku and no. 880 in Merimasku are brownish. The minerals in both are the same: quartz, plagioclase (An_{22-24}), microcline (0 to 5 per cent of the total feldspar), biotite, garnet, magnetite, and zircon. The presence of garnet is of special interest. As a rule the main part of the rock contains only a few small garnets, while in the pegmatitic portions they are larger and more abundant. The iron seems to have enriched in the rest magma and formed garnet especially in the pegmatitic stage. The question arises as to how the trondhjemitic magma became enriched in the elements which are necessary for the formation of garnet. Probably the magma has assimilated the older rocks, which contained garnet. But these older rocks also contained cordierite, which should also be found in the trondhjemites. However, only one small, partly altered cordierite grain was found in the thin sections investigated. This cordierite occurs in the bluish-gray trondhjemite in the middle part of Luonn-

onmaa (Fig. 15). The southernmost part of the charnockite pluton consists here of a light-brownish-gray rock, similar to no. 891. The bluish-gray trondhjemites are separated from it by a narrow layer of cordierite-garnet kinzigite, which contains in addition to the common granitic veins also trondhjemitic veins. The brownish border

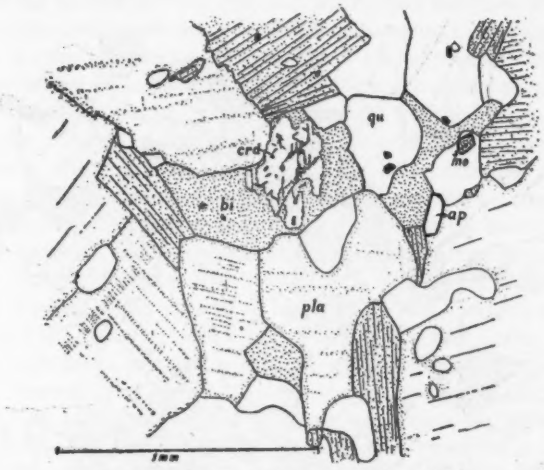


FIGURE 15.—*Cordierite changing into biotite*
Trondhjemite, Luonnonmaa (no. 900, Pl. 1). (pla) plagioclase; (qu) quartz; (bi) biotite; (crd) cordierite; (mo) monazite; (ap) apatite.

facies contains about 10 per cent microcline and no garnet, while the bluish-gray trondhjemite contains microcline only as antiperthite, and garnet in addition to the common minerals (Fig. 16). Trondhjemite contains angular and oblong dark inclusions similar to the basic inclusions in the diorite-trondhjemites of the Kalanti area. Contact zones grade into the kinzigite with trondhjemite veinlets. Garnet is abundant especially near the contact zones originating undoubtedly in the kinzigite. But cordierite, which is also common in the kinzigite, was found in the trondhjemite only in one case, and then it was changing into biotite (Fig. 15). The cordierite has probably reacted with the magma forming biotite whereas the garnet has been stable and recrystallized. The reaction cordierite → biotite requires a certain amount of potash, and that may account for the lack of potash feldspar in these narrow trondhjemite bodies which occur in contact with the microcline-bearing intermediate charnockites. The reaction may be expressed as follows:

$$2(Fe, Mg)O\cdot 2Al_2O_3\cdot 5SiO_2 + 2 \times K_2O\cdot Al_2O_3\cdot 6SiO_2 + 12 \times (Mg, Fe)O\cdot SiO_2 + 4H_2O \rightarrow$$
cordierite + 2 microcline + 12 hypersthene + 4 water
$$4H_2O\cdot 2K_2O\cdot 11(Fe, Mg)O\cdot 3Al_2O_3\cdot 11SiO_2 + 3(Fe, Mg)O\cdot Al_2O_3\cdot 3SiO_2 + 15SiO_2$$
2 biotite + almandite + 15quartz

This assumption is supported by the fact that associated with the charnockites are acidic members with some potash feldspar and no garnet—*e.g.*, nos. 891 and 905. These granite-trondhjemites could well represent the primary magma which invaded the kinzigites, reacted with them, and formed garnet-bearing trondhjemites.

The narrow trondhjemite lenses east and west of Merimasku church and no. 880 consist of a medium-grained brownish-gray rock, which in places grades into a light-bluish-gray type. Garnet is common in both varieties. The composition of plagioclase is An_{28}. The wider trondhjemite mass south of Turku most probably belongs

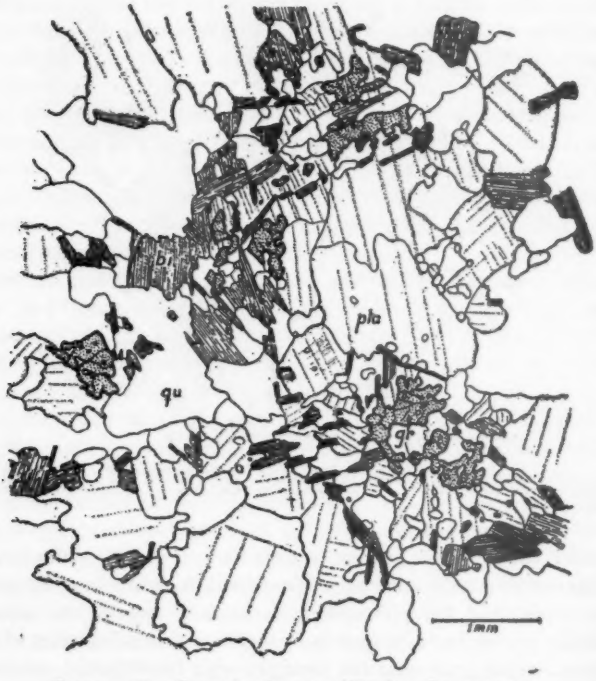


FIGURE 16.—*Garnet-bearing trondhjemite, Luonnonmaa* (No. 900, Pl. 1)

to this group. Like the other trondhjemites associated with the charnockites this rock also contains some small garnets, is coarse- or medium-grained, shows no cleavability, and is thus harder to split with the hammer than the normal trondhjemites of the Kalanti region.

In the acidic members of the charnockite series the pyroxenes occur only rarely. Usually the biotite is the only dark constituent, and garnet may occur with it. Monoclinic pyroxene was found in a light-gray fine-grained rock 1 km. east of Lenklahti, Rymättylä. This pyroxene trondhjemite strongly resembles the light-gray varieties of the pyroxene trondhjemites of Putsaari.

Hypersthene, badly altered into biotite, is common in the coarse- and medium-grained brown or dark-brownish-gray rocks 6 km. northeast of Turku. The amount of potash feldspar varies locally: in the western part of this area the rock is trondhjemitic, while in the eastern part the potash feldspar may be abundant especially in the

pegmatitic portions. The hypersthene trondhjemites and granites grade into light-gray or reddish-gray rather inhomogeneous rocks which contain potash feldspar and garnets in varying amounts. Because of the presence of garnets these rocks resemble the Kakola granite, but in contrast to it these garnet-bearing granites contain no cordierite, and the garnets are very unevenly distributed, and their size varies greatly. Accumulations occur ranging from 3 to 6 cm. in diameter, and individual grains range from 1 to 15 mm. in diameter. Furthermore, the pegmatitic differentiates which occur in stocks and veins make the inhomogeneity still more conspicuous.

The garnet-bearing granite in the west-east zone through Satava is very similar to these granites. The rock in Satava is also rather inhomogeneous, contains remnants of older rocks, kinzigites, and also amphibolites. The pegmatitic differentiates are common, and the garnets are of varying size and scattered. Also the amount of biotite varies greatly. These granites are associated with the hornblende- and pyroxene-bearing brownish and grayish granite-trondhjemites, in which the percentage of the potash feldspar ranges from 20 to 40 per cent of the total feldspar, and which in places may contain small garnets. The structure and the color of these granite-trondhjemites are typical of the rocks of the charnockite series. Hypersthene is absent, but light-brown aggregates of biotite, obviously alteration products of hypersthene, are common (Pl. 4, fig. 1). Thus probably the garnet-bearing granites south and east of Turku belong geologically to the charnockite series, representing its most acidic member. The garnets may originate from the kinzigites in a manner similar to that described later in the section "Kakola granite." The differences between true Kakola granites and garnet-bearing granites may be attributed to differing stages of granitization as discussed below.

Pegmatites are trondhjemitic or granitic. The former consist of plagioclase, quartz, and some biotite, whereas the latter contain abundant microcline in addition. The trondhjemitic pegmatites occur as veins in trondhjemites or in the kinzigites near the contacts. In the garnet-bearing trondhjemite lenses of Merimasku-Naantali zone these pegmatites contain abundant idiomorphic garnets.

Pöytyä diorite and associated rocks.—In the eastern portion of the parish of Masku, in Vahto, and in Paattinen vast areas are occupied by light-bluish-gray to gray rocks whose composition ranges from trondhjemites to granite-trondhjemites and to hornblende trondhjemites. Potash feldspar occurs locally, and most of it is due to migmatization. Usually the biotite is the only dark constituent, but hornblende may also occur in varying amounts. Plagioclase (An_{22-30}) and quartz are the light constituents. Toward the west these rocks grade into charnockites which belong to the same series. East of the present area this type is the most common and occurs locally with the volcanics causing their partial granitization. In this region trondhjemitic and hornblende trondhjemitic varieties prevail also, while rocks containing potash feldspar are rare. In the parish of Pöytyä a more basic and coarse-grained variety occupies large areas. This so-called Pöytä diorite is dark gray and shows a well-developed orientation of hornblende and biotite. Plagioclase (An_{26-38}) and quartz are the light minerals, zircon and magnetite the accessories. Potash feldspar may occur in small amounts and is probably secondary. Epidote is sometimes abundant as large grains associated with the dark constituents. The plagioclase then

contains a lower percentage of anorthite (An_{22}). Epidote shows $\beta = 1.755 \pm 0.001$ and contains 26 per cent Fe component (Winchell, 1933). Epidote is common also in the hornblende trondhjemites and diorites west of Pöytyä as for instance in the gray medium-grained schistose diorite 4 km north of Paattinen (no. 1220). This

TABLE 5.—*Diorite, Pöytyä*

Chemical composition		Minerals		Norms		Niggli values	
SiO_2	66.64	qu	28.61	qu	25.90	si	261.5
TiO_2	0.66	pla(An_{34})	42.60	ab	29.25	ti	0.19
Al_2O_3	14.81	mi	4.80	an	19.78	al	34.45
Fe_2O_3	1.57	bi	11.20	or	10.17	fm	26.35
FeO	3.16	ho	7.54	wo	2.09	c	21.60
MnO	0.08	ep	3.63	en	4.64	alk	17.60
MgO	1.87	orth.	0.20	hy	3.58	qz	91.1
CaO	5.09	ap	0.19	ap	0.19	k	0.25
Na_2O	3.44	ti	0.73	il	1.25	mg	0.4
K_2O	1.74	mt	0.30	mt	2.28	c/fm	0.82
P_2O_5	0.08			H_2O	0.68		
H_2O+	0.48						
H_2O-	0.20						
Total.....	99.82		99.80		99.81		

diorite is closely associated with the lighter-gray hornblende-bearing diorite trondhjemites and granite-trondhjemites of the same area. A few orthite grains were found with epidote in some thin sections.

The association epidote-plagioclase (An_{22-30}) would indicate pressure and temperature conditions of an epidote amphibolite facies during the crystallization as was also pointed out by Ramberg (1943; 1945).

Table 5 shows the chemical and mineralogical composition of the diorite 3 km south of the Pöytyä church.

The chemical analysis shows high soda and lime and low potash content, thus the characteristics typical of the trondhjemite and charnockite series. The exceptionally high percentage of CaO accounts for the occurrence of abundant epidote. At the same time the content of Al_2O_3 is lower. The amounts of the other oxides are the same as in the intermediate charnockites.

INTRUSIVES OF THE SECOND CYCLE

Rocks of the granite series.—The intrusives just north of Turku seem to belong to a differentiation series whose most acidic member is a red granite rich in microcline. The various members are gabbro, diorite, granodiorite, and granite which penetrate the kinzigites and the amphibolites as small concordant oblong bodies. The coarse-grained pegmatitic portions of the red granite may cut the other rocks discordantly. In addition a red granite containing garnet and cordierite occurs about 3 km north of Turku on the road Kuninkaantie (no. 1168). Integration-stage analyses made from various members of the differentiation series occurring north of Räntämäki are shown in Table 1.

Gabbro (no. 14 in Table 1) is a coarse-grained and unoriented dark rock in which the short hornblende prisms show glittering cleavage surfaces. Plagioclase (An_{42-50}) shows albite, pericline, and Carlsbad twinnings. Zoning and partly idiomorphic habits are common. The hornblende has $\gamma = 1.676 \pm 0.001$ dark green, $\beta = 1.672 \pm 0.001$ brownish green, $\alpha = 1.656 \pm 0.001$ greenish yellow, $\gamma \wedge c = 16.5°$. It carries brown patches of biotite and scales of ilmenite parallel to the cleavage. The biotite with $\gamma = 1.640 \pm 0.001$ is the common brown variety. Roundish magnetite grains are abundant and occur commonly with the hornblende. Idiomorphic apatite crystals are also common. Some small quartz grains are seen between the feldspar crystals. There are no cleavage or parting planes, but the rock has been quarried and used for tombstones because it is dark and surfaces take a good polish.

Diorite (Table 1, no. 15) is somewhat lighter-colored and usually more brownish than the gabbro. The cleavage surfaces appear always dull. Reddish-brown, strongly pleochroic biotite forms the bulk of the dark minerals, and the amount of hornblende is rather small. Ore is also abundant, and in addition sphene occurs in large amounts. It is usually associated with the ore and suggests that the ore may be ilmenite-magnetite. Microcline occurs only as angular antiperthitic inclusions in plagioclase.

The diorite grades into a lighter and somewhat fine-grained rock along the margins and in broad veins, which penetrate the surrounding kinzigites. The measured mineral composition and the calculated chemical analysis of this lighter rock are shown in Table 1 (no. 16) and indicate granodiorite. Biotite is the only dark constituent, the percentages of quartz and microcline are larger than in the diorite, and the plagioclase is An_{30}. The amounts of ore and sphene are still considerable. Near the contact zones of this rock the kinzigite contains broad veins of a still lighter rock which contains some garnet. The garnets occur only along the borders of the small masses. The veins were probably formed as an assimilation product of the granodioritic magma and the kinzigites. Their composition is most likely near that of the so-called Kakola granite. In addition light-gray Kakola granite with abundant garnet occurs in the same area as veins and small masses in the kinzigite.

A small granodiorite intrusion on the northern shore of the island Taipalus belongs to this group. It cuts the veins of adjacent kinzigite migmatite discordantly (Fig. 17).

Red microcline granites occur with the gabbros and diorites north and south of Turku and in narrow lenses in the kinzigites and amphibolites in the islands Ruissalo and Hirvensalo west of the same city. They seem to be closely related to the basic rocks described above. The contact between granite and the basic rocks appears sharper than that between the various basic members. The grain size varies greatly, and as a rule the large masses are coarse-grained, while the narrow long lenses are fine-grained. Potash feldspar constitutes about 56 per cent of the total feldspar. It occurs usually as thick laths which are Carlsbad twins showing a typical microcline "gridiron" structure only in a few patches, while other grains are regularly twinned. Plagioclase (An_{21}) is partly idiomorphic and twinned, according to the albite law. The amount of biotite is rather small. The calculated analyses of the granite no. 1159 is shown in Table 1 (no. 17).

The red coarse-grained pegmatites rich in microcline are common in this rock series. They invade the oldest complex and the intrusives concordantly as well as discordantly. Quartz and plagioclase occur together with the microcline. Graphic granite and perthitic intergrowth of potash and soda feldspar are common. Biotite is generally lacking. Cordierite and garnet occur in some red pegmatitic differentiates in the

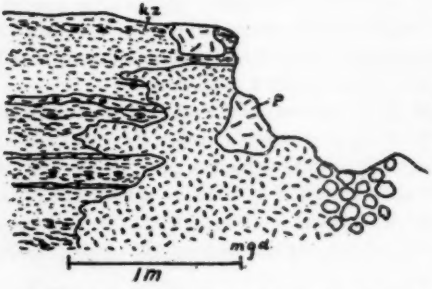


FIGURE 17.—*Granodiorite (mgd) cutting the migmatitic veins discordantly*

Younger pegmatite (p) transects both. Kinzigite (kz) is oldest. Northern shore of the island Taipalus.

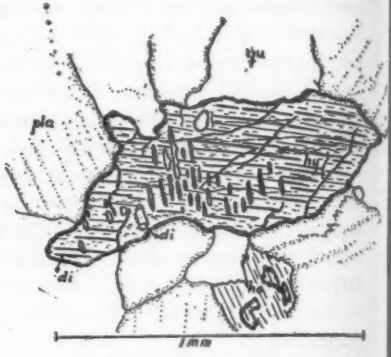


FIGURE 18.—*Diopside in lamellae parallel to (010) and dome planes in hypersthene*

Anorthosite of Kolinummi, Yläne (no. 957, Pl. 1).

kinzigites. Some hornblende may occur in the pegmatites which penetrate the volcanic rocks—*e.g.*, 5 km east of the Rusko church.

POST-TECTONIC INTRUSIVES

Diabase dikes.—Diabase dikes cut the rocks of the first and second cycle discordantly and are to be correlated with the diabase dikes of the Kalanti area (Hietanen, 1943, p. 63). They are numerous but rather small; for instance in the southern coast of the island Salomaa most of them range only from 1 cm. to ½ m. in width and do not continue very far from the coast. Only one longer dike, parallel to the foliation of the kinzigites, occurs in the southern part of Rymättylä. This dike was reported by Moberg (1890). It is 1½ to 2 m. thick and about 6 km. long. According to W. Ramsay it is fine-grained and consists of augite, bronzite, basic plagioclase, and some biotite, calcite, and quartz.

Anorthosite.—The oblong anorthosite body in Kolinummi parallels the general strike and the contact of the Laitila rapakivi intrusions and is most likely associated with them. It consists mainly of coarse-grained bluish-gray labradorite (An_{45-49}). Some tremolite ($\gamma = 1.640 \pm 0.002, \alpha = 1.622 \pm 0.002$) occurs as a dark constituent. In the southern part of the area the rock is medium-grained, and its structure is typical of diabase. It contains in addition to labradorite and hypersthene some diopside, biotite, and tremolite. The plagioclase shows Carlsbad, albite, Ala, pericline, and complex twinning. Diopside occurs in small grains and also as lamellae in some hypersthene grains (Fig. 18). Microcline and myrmekite are also present. Magnetite and apatite are the common accessories.

The hypersthene shows a peculiar structure which was found also in some norites and charnockites near Turku (Pl. 4, fig. 2). It contains thin lamellae of some mineral, which has lower indices of refraction than the hypersthene, a low or moderate birefringence, and inclined extinction. The lamellae parallel the parting plane (010) in

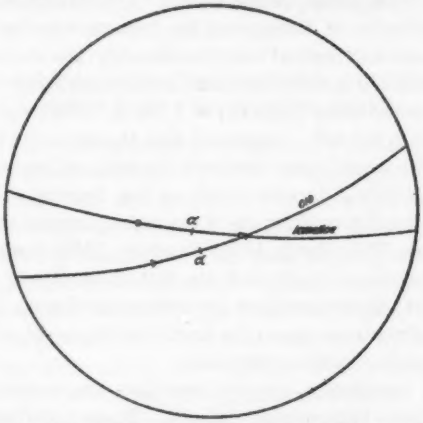


FIGURE 19.—*Diagram showing orientation of the optic axes and acial planes of a hypersthene and the lamellae*

Pyroxene amphibolite 6 km southeast of Turku (no. 1336, Pl. 1).

the hypersthene and are in most cases so thin that the determination of further optical properties is impossible. For closer investigation several thin sections were made from the same specimen, whose pyroxenes show rather broad lamellae. However, only in three grains were they broad enough for the determination of the optic axes and their orientation. In two cases the main mineral and the lamellae show negative optic angles of about the same value: $-2V = 53°$ in one grain, and $-2V = 80°$ in another grain. In the former grain the axial plane makes an angle of 16° and in the latter 30° with the (010) face of the main mineral. A hypersthene from a pyroxene amphibolite no. 1336 6 km southeast of Turku shows similar lamellae with a weak birefringence (Pl. 4, fig. 3). In this case the optic plane of the lamellae makes an angle of 24° with the (010) plane of the hypersthene. At the same time one of the optic axes of both minerals coincides (Fig. 19). The optic angle of the hypersthene is $-2V = 50°$, while that in the lamellae was measured as $-2V = 53.5°$. The difference in the values of the optic angles may be due to errors in measuring. The indices of refraction do not seem to differ in the lamellae and the hypersthene. Probably therefore these lamellae are also hypersthene—the (010) face of the main mineral and (041) plane of the lamellae have grown together. In the third grain the thin lamellae parallel to (010) are probably diopside. They have a lower refractive index and higher birefringence than the hypersthene. The extinction angle is 40°, and the (100) face of the diopside has grown together with the (010) face of the hypersthene. In addition diopside lamellae parallel to the base and dome planes are common in all hypersthenes in the anorthosite of Kolinummi.

A. H. Phillips and H. H. Hess have described diopside lamellae in hypersthene and explained them as due to exsolution after the crystallization of a hypersthene containing a certain amount of CaO. In the Bushveld type and in the pyroxenes from Stillwater these lamellae parallel (010) (Hess and Phillips, 1938). In the latter location (100) faces and dome planes are also known. The extinction angle of the lamellae is 40°. In the anorthosite of Kolinummi the refractive indices of the hypersthene measured by the immersion method vary considerably: the lowest α is 1.719 ± 0.001, while the highest γ is 1.750 ± 0.001, but the birefringence is only 0.011 ± 0.002. The diopside shows somewhat lower indices ($\gamma = 1.734 \pm 0.002$), and $\gamma - \alpha$ is larger than the birefringence which is 0.022. Supposing that the refractive indices vary with the ratio [FeO]:[MgO] this would mean that both rhombic and monoclinic minerals form a series in which [FeO]:[MgO] varies within certain degrees. In the light of the investigations concerning the equilibrium of the orthopyroxene and the clinopyroxene (Bowen and Schairer, 1935; Barth, 1936; Ramberg, 1945) the intimate intergrowth of diopside and hypersthene would indicate that the rock was crystallized near the transformation point clinohypersthene $\rightleftarrows$ orthohypersthene. The calcium-bearing compound separated contemporaneously from the pigeonitic solution as a diopside and remained as lamellae in the hypersthene.

The Kolinummi anorthosite greatly resembles the anorthosites described by Sederholm (1934) from Höggrunden, Eckerö. These anorthosites occur with the so-called ossipite-diabases, which are considered closely associated with the rapakivi, but older. The pyroxene in these diabases consists of enstatite-augite according to Sederholm. The light-colored coarse anorthosite at Höggrunden consists of almost pure labradorite; its color, however, is light tan.

Rapakivi.—The present paper does not include a study of rapakivi. Only points of general interest are briefly described. The rapakivi is the youngest granite in the area and clearly post-tectonic. The contacts are discordant, and angular inclusions of the country rocks occur along the marginal zones, for instance in the easternmost part of the Vehmaa rapakivi (the westernmost rapakivi area in Pl. 1). This easternmost part consists of a medium- and even-grained red type, so-called Vehmaa granite, while the area south of it—the area which penetrates into the older complex like a wide bay (Pl. 1)—consists of a red fine-grained porphyritic rock. The southern part as well as the southern part of the other large rapakivi pluton in the northernmost part of the map (Pl. 1), so-called Laitila rapakivi, consist of a normal coarse-grained variety which contains huge rounded orthoclases rimmed by plagioclase. The small rapakivi 4 km south of Mynämäki consists of the same type, whereas the other small occurrence 3 km to the north consists of the red fine-grained porphyritic variety.

Quartz porphyry dike.—A discordant dike of red fine-grained quartz porphyry cuts the older rocks on the southern shore of the southermost peninsula of Ruissalo. The oldest rocks here consist of kinzigite and intercalated amphibolite and leptite layers. The microcline granite penetrates them in the upper part of the slope. The quartz porphyry cuts all rocks striking N. 60° W. with vertical contacts. Near the shore line it is about 6 m. thick, thins upward, and ends in the upper part of the slope.

The center of the dike is porphyritic, while the borders are fine-grained. The minerals in both are quartz, feldspar, muscovite, biotite, chlorite, and magnetite. Some quartz and potash feldspar occur as partly idiomorphic phenocrysts 1 to 10 mm. in diameter (Pl. 4, fig. 4). Intergrowth of quartz and feldspar is common in the groundmass especially adjacent to the quartz phenocrysts. Feldspar contains abundant fine sericite and ore inclusions.

CHEMICAL AND MINERALOGICAL COMPOSITION OF THE ROCKS OF THE CHARNOCKITE AND GRANITE SERIES

The main oxides of the chemical analysis from 1 to 17 were plotted into the variation diagrams (Fig. 20). The charnockite series is characterized by low potash and ferrioxide and high soda, lime, and magnesia contents. These properties bring it chemically near the trondhjemite series of the Kalanti district. The rocks of the granite series differ greatly from both. The most striking difference between the granites and charnockites is the enrichment of potash in the acidic members of the granites. In the charnockite series the lime dominates slightly over the ferro-oxide and both over the magnesia. In the granite series the differences between the amounts of these three oxides are far more striking: FeO is rather high, while MgO is much lower than in the charnockite and trondhjemite series. Lime is also lower. Thus in the granites iron conspicuously dominates the lime and especially the magnesia.

The chemical composition of the charnockites is very similar to that of the trondhjemites. Moreover, the bulk of the most acidic charnockites are identical with the trondhjemites not only chemically but also mineralogically.

The difference between these two series is in the mineralogy of the basic and intermediate members. The charnockite series contains hypersthene and diopside, but the corresponding members of the trondhjemite series contain exclusively hornblende and biotite, or only biotite, as femic minerals. However, in the Kalanti area pyroxene-bearing members are associated with the trondhjemites (the pyroxene trondhjemites of Putsaari and Mynämäki), and in the Turku area hornblendites, gabbros, and diorites, in which the pyroxenes are lacking or scarce, grade into charnockites. There seems to be no age difference between these two types. Rather the presence or lack of pyroxenes is local, and in some cases the lack is due to alteration. Remnants of diopside in the hornblende-bearing rocks especially are common. Aggregates of light-brown biotite are found in several diorites and suggest that the hypersthene is primary. For instance, hornblendite 1140 and gabbro 1142 grade into a diorite which contains such biotite, and then into a charnockite, in which the hypersthene is only partly altered.

The Pöytyä diorite series is mineralogically as well as chemically very similar to the trondhjemite series. Soda and lime content is high, and potash low. Epidote is locally abundant and associated with the femic constituents. Plagioclase contains less anorthite. Aside from the epidote and the somewhat greater variation in the amount of potash feldspar the mineral compositions of the two series are identical. Their structures and mode of occurrence in the field are also alike. The various members of all three differentiation series invaded the older rocks during the same

period of folding, the charnockites perhaps somewhat later than the others. This assumption is based on the structural features, and the mineral development supports the view. In the trondhjemite series it is the same as described from the Kalanti

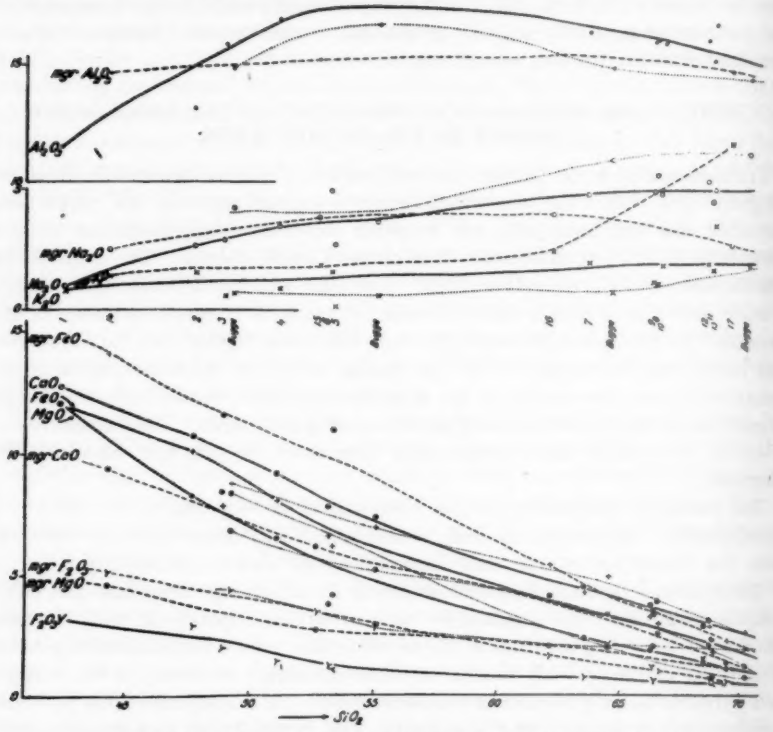


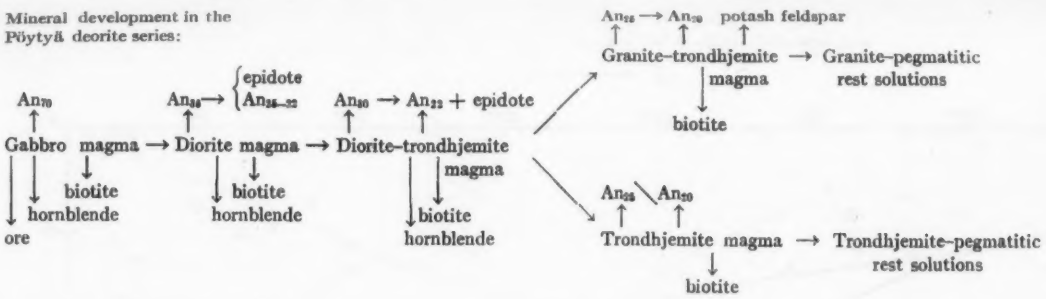
FIGURE 20.—*Main oxides of the rocks of charnockite and granite series*

district (Hietanen, 1943, p. 58). A corresponding scheme for the Pöytyä diorite and associated rocks would be as shown in the accompanying diagram.

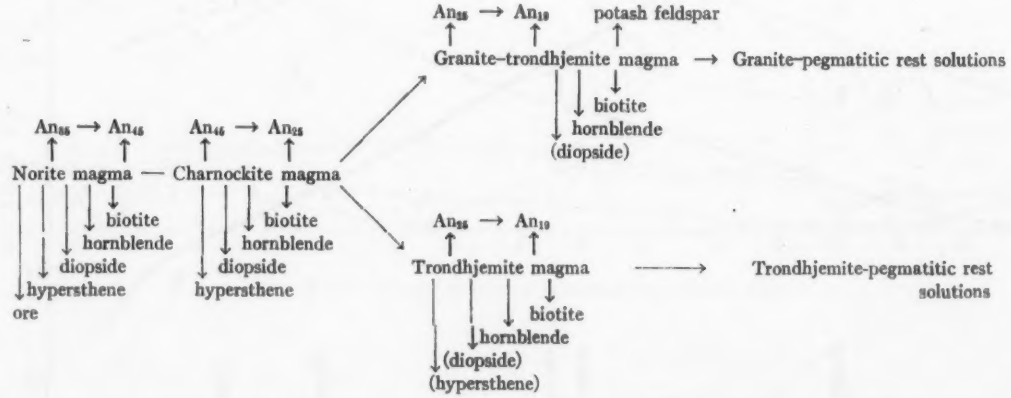
For further comparison of the chemical composition of the charnockite and trondhjemite series the Niggli-values al, fm, c, and alk for analyses 1 to 13 were plotted into the diagram (Fig. 21), which illustrates the equivalent values of the Kalanti trondhjemites (Hietanen, 1943, Fig. 27). In Figure 21 the Niggli values of the charnockites are near those of the trondhjemites. The charnockites 8, 9, and 10 are more acidic than the hornblende trondhjemites, their si values are between those of the hornblende trondhjemites and the biotite-bearing diorite-trondhjemite, while the granite-trondhjemite 12 and the trondhjemite 11 show approximately the same si values as the biotite-bearing diorite-trondhjemite. In the field the granite-trondhjemite 13 looks like the fine-grained trondhjemite of Haidus. No. 7 resembles the hornblende trondhjemites. No. 6, which has the same minerals

Mineral development in the Pöytyä deorite series:
An₂₅ → An₀ potash feldspar

Mineral development in the Pöytyä deorite series:



and in the charnockite series:



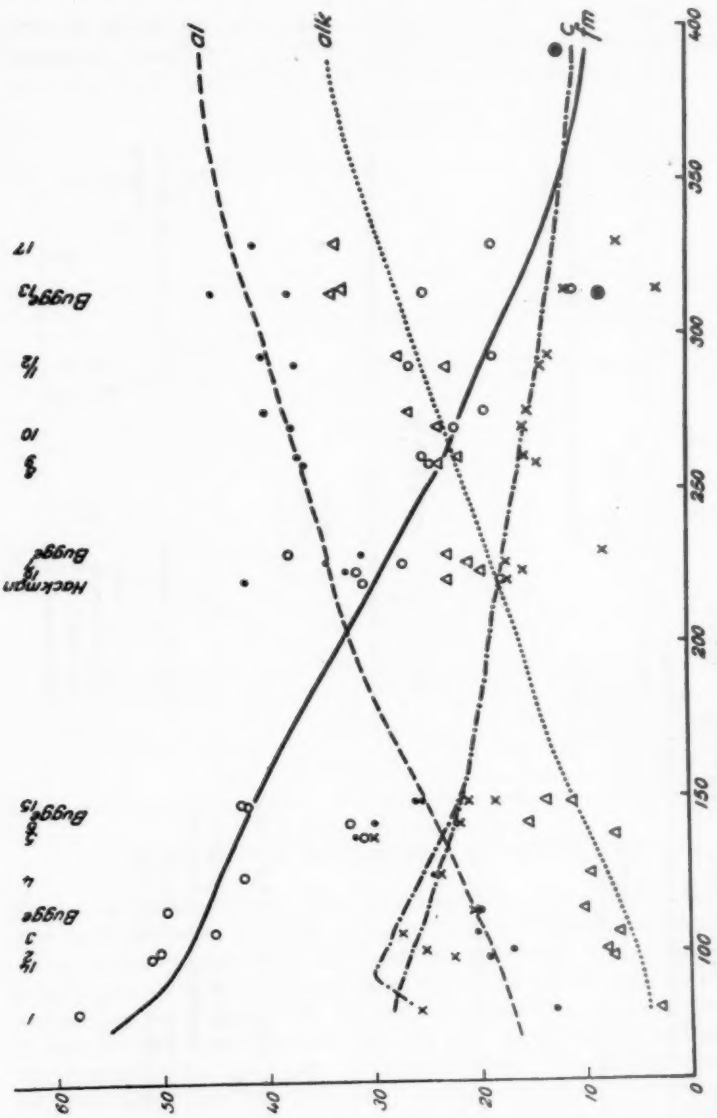


FIGURE 21.—*Niggli values of the rocks of the charnockite and granite series*

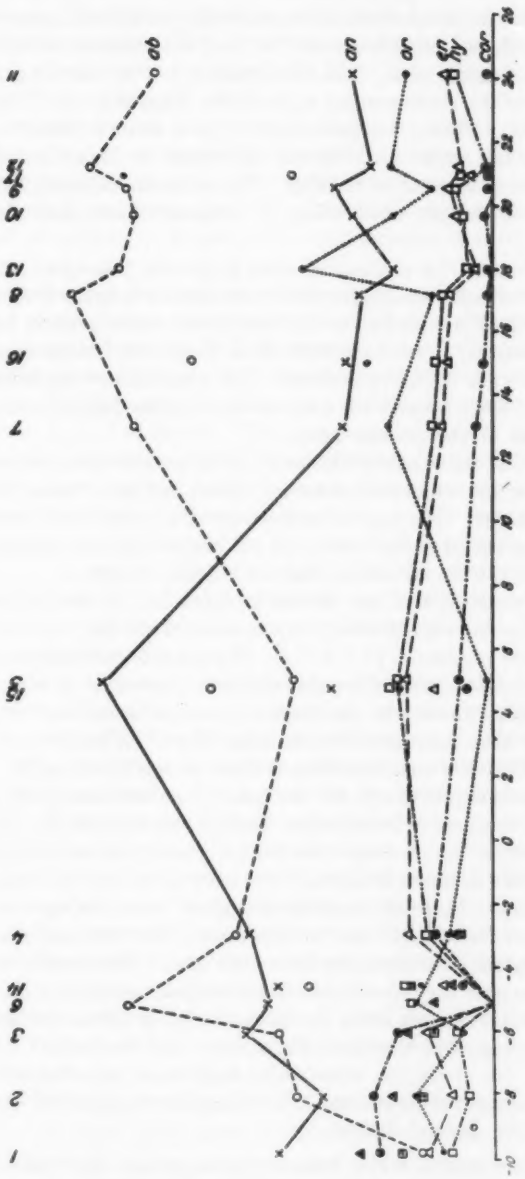


FIGURE 22.—*Norms of the rocks of the charnockite and granite series*

FIGURE 21.—*Niggli values of the rocks of the charnockite and granite series*

as no. 7, is a diopside-bearing diorite lens in the hornblende charnockite pluton. It is abnormally low in quartz which makes its si value smaller than one would expect according to the al, fm, and alk values. It is an accumulation of ferromagnesian minerals in a more acidic rock. The hornblendite (no. 1) and the gabbro (no. 3) are very similar to the corresponding rocks of the Kalanti area. The basic norite (no. 2) is gabbroic, while no. 4 is more acidic. Its si value is between the gabbros and diorites. In the norite no. 5 fm and alk values are lower, and the al and c values higher than in the normal norities. This is shown mineralogically in a high percentage of calcic feldspar, which brings its composition near that of the anorthositic norities.

The Niggli values for the charnockite from Kakskerta (Hackman, 1923) and the arendalite from Norway (Jens Bugge, 1940) were inserted in figure 21 for comparison. The si values of both are similar to the hornblende trondhjemites, but the other values differ somewhat. The charnockite from Kakskerta has considerably higher al and alk values, while the fm value is low. The arendalite on the other hand has a very low c value, while fm and alk are somewhat higher and al lower than in the charnockite series of the present area.

The al, fm, c, and alk values of the rocks of the granite series were also plotted in figure 21. The alk values are somewhat higher, and the c values lower than in the charnockite series. The mg values are very low in the basic members. The k values are approximately the same as in the normal gabbros and diorites. The granite no. 17 is rich in microcline, and its k value is high.

The norms of analyses 1–17 are plotted in figure 22. In the intermediate and acidic charnockites the soda distinctly dominates potash; the ratio ab:or is about 4:1, whereas in the granite no. 17 it is 1:2. The granite-trondhjemite no. 13 shows ab:or = 2:1. In the charnockites the normative anorthite is somewhat higher than the orthoclase, whereas in the granite-trondhjemite and in the granite it is lower; en and hy have approximately the same values. A small excess of alumina occurs in the most acidic members; lime in them thus is wholly salic. Also in the intermediate members practically all lime goes to the anorthite molecule, whereas in the basic members part of it enters the diopside and hornblende. In the norites 2 and 4 the values for ab are larger than for an, whereas in the hornblendite 1 and gabbro 3 and 14 the an value is higher. The amounts of wo, fo, and fa increase in the basic members. Magnesia dominates slightly over the ferro-oxides in the hornblendite no. 1 and gabbro no. 3. However, if the iron oxides are reckoned as FeO the magnesia is consistently below the iron. The normative magnesium silicate dominates over the ferrosilicates in the normal members of the charnockite series, whereas in the granite series the amount of iron silicates is always higher than that of the magnesium silicates. The norite 5 and the diorite 6 differ from the other members. No. 5 has an abnormally high value for normative anorthite, due to the large amount of plagioclase. No. 6 has abnormally low silica and a higher ab value than the normal dioritès.

COMPARISON WITH THE OTHER REGIONS AND THE ORIGIN OF THE CHARNOCKITES

Washington (1916) made a series of new analyses of the various members of the charnockite series of Madras, India, originally described by Holland (1900). He

also compares this occurrence to several other regions which have the main features in common—*e.g.*, Ekersund, Soggendahl, and Bergen district in western and southern Norway (Kolderup, 1896; 1903), Ellesmere Land and Prudhoe Land (C. Bugge, 1910), the Cortlandt series near Peekskill, New York (Rogers, 1911), and Ivory Coast, West Africa (Lacroix, 1910).

The original charnockites in southern India form a series from basic hornblende hypersthenites through intermediate hypersthene quartz diorites to acidic hypersthene granites (charnockites). Charnockites in Ceylon are, according to Adams (1929), very similar to those in India. The rocks show a tendency to parallel arrangement of minerals. Hypersthene and diopside are the prominent dark constituents.

In the present area there are several members, which greatly resemble these charnockites, especially the norites and the intermediate charnockites. Hypersthene and abundant plagioclase (Ab_3An_1) characterize these rocks. A notable amount of lime is always present also in the intermediate and acidic members, but not more than enough to form the anorthite molecule. In the basic members part of it enters the diopside and the hornblende. The percentage of K_2O in the acidic members of the charnockite described earlier is larger than that of the Na_2O. The normative amounts of soda and potash are equal, whereas in the present area the soda dominates greatly over the potash also in the acidic members. In this respect the present rocks greatly resemble the charnockites of the Arendal district described by J. Bugge (1940). However, there are certain differences as shown in the variation diagram (fig. 20) and in figures 21 and 22. In the arendalites the domination of Na_2O over K_2O is still greater, the percentage of CaO lower, and that of Fe_2O_3 higher than in the charnockites under discussion.

Each area seems to have its individual characteristics which may be attributed to differences in differentiation or to reactions between the older rocks and the magma. Also the rocks with which the charnockites are associated differ for every region. In Norway they are labradorite or andesine anorthosites and ilmenitites. In India hypersthene-bearing pyroxenites are the most basic members. In the Cortlandt series a mica granite without hypersthene occurs, but its relations with the charnockites are uncertain. In the present area the intermediate charnockites grade into trondhjemites and granite-trondhjemites in which biotite is often the only dark constituent, but pyroxene-bearing acidic members also occur. In addition to hypersthene the present series contains also diopside, and among the intermediate and basic members there are hypersthene-free varieties, but monoclinic pyroxene and hornblende or only hornblende abound.

The charnockite series of Uganda (Groves, 1935) resembles the present area in this respect. The diopside always occurs with hypersthene, and the hornblende is the third essential dark constituent.

The earlier investigators considered the charnockites magmatic, but in 1935 Groves made their origin questionable by describing the Uganda series to be metamorphic. Parras (1941) discovered in southern Finland pyroxene gneisses which grade into calcareous leptites and are clearly metamorphic. In his latest papers J. Bugge (1943; 1945) discusses the origin of the arendalites in the light of the diffusion in the solid state and concludes that these rocks are migmatic, having been

formed by metasomatic transformation of the pre-existing rock complex. Arendalites would thus be comparable to the charnockites of Uganda. Both Groves and Bugge could prove the reaction series in which hypersthene was derived from biotite or diopside, and the diopside from hornblende. The charnockites of the Turku area do not show this kind of reaction but only normal alterations common in igneous rocks. As a rule the dark minerals in the unaltered rocks show relations which refer to a contemporaneous crystallization. The sequence of crystallization of monoclinic and orthorhombic pyroxenes varies. In some thin sections the hypersthene grains are surrounded by diopside (Fig. 10), while some other sections show just the opposite relations. Relations between hypersthene and biotite, as well as those between diopside and hornblende, are like those common in the igneous rocks.

The existence of all members of a complete series from the basic norites to hypersthene trondhjemites, their mode of occurrence in the field, and the structures suggest a magmatic rather than metamorphic origin. The occurrence of the trondhjemites in association with the charnockites seems difficult to explain from a magmatic point of view. The only difference between these two series is a crystallization of pyroxenes in the charnockites instead of hornblende and biotite. In the metamorphic field this would mean higher temperature-pressure conditions in the charnockites. In the magmatic differentiation it would be a result of lower content of water. The common grading from one series to another around Turku indicates close relations between these two. The alterations indicate the general tendency to be from a higher to lower temperature-pressure field. The reaction hypersthene → biotite seems to explain the lack of hypersthene in many cases. The texture under the microscope shows that in most cases this alteration is due to a later hydrothermal activity and to an introduction of potash. Further, the same reactions may well have occurred when the dry magma assimilated watery sediments. Only the grading of sedimentary amphibolites into norites suggests metamorphic origin of charnockites. The bulk of the charnockites show igneous features. In contrast to the arendalites they cannot be formed through reactions in a solid state, but rather the increased temperature-pressure conditions caused the formation of a palingenetic magma.

The formation of the charnockite magma may be compared to that of cordierite-garnet granites described below. Both have a smilar granoblastic texture differing from the hypidiomorphic texture of common igneous rocks. The cordierite-garnet granites are clearly formed by the co-operation of metasomatic, anatectic, and palingenetic processes. The amount of material introduced by metasomatism is calculated for the cordierite-garnet granites (Table 7). In the charnockites it may be still greater, and the role of older rocks may be less important since the amphibolites form the minor portion of the older rocks. The high content of soda, lime, and magnesia in the charnockite series is probably due not only to the composition of the assimilated sediments but also to the original composition of the upward-migrating ichor. The importance of the composition of a granitizing magma or solution is clearly proven in the composition of younger cordierite-garnet granites, in which case the solutions came from red microcline granites. The end product in the granitization shows the same features which are characteristic of granitizing magma. All three series—charnockites, trondhjemites, and Pöytyä diorite—are closely associated, and their

structures prove them to be syntectonic. The differences in their mineral assemblage are obviously due to the amount of water in the older rocks and to the temperature-pressure conditions during their emplacement or formation. Probably there was a magma chamber in the Turku area below the present earth level. This magma provided heat and ichor to the older rocks. Those nearest the magma chamber received more heat and metamorphosed in a higher facies releasing part of their water. The continued rising of magma was accompanied by its differentiation and assimilation of older rocks. The palingenetic, anatectic, and metasomatic processes gave rise to the great variation of the rocks in the surroundings of the original magma chamber. Those parts of the magma which wandered further from the original magma chamber gave rise to the formation of the trondhjemites and Pöytyä diorite. They released a part of their heat and assimilated watery sediments. But trondhjemites were locally formed also in the higher temperature field as a result of reactions between the older rocks and rising ichor. Equilibrium was attained during metamorphism since the metamorphic rocks show the same mineral assemblage as the igneous ones in each area.

STRUCTURES

STRUCTURE OF THE SEDIMENTARY AND VOLCANIC ROCKS

Bedding or layering can be recognized with certainty only in the best-preserved areas. Intense migmatization has generally obliterated primary structures. However, its traces may be recognized in the migmatized outcrops by a careful observer. In the kinzigites it is visible similarly, as described from Kalanti (Hietanen, 1943), in the alternation of coarse-grained layers rich in biotite, cordierite, and often also garnet with the finer-grained, lighter-colored layers, which contain usually more quartz and less biotite. Cordierite is usually lacking, but garnet is common. The thickness of the layers varies greatly. In the normal kinzigites the coarse-grained layers rich in cordierite are usually 1 to 5 m. thick, while the fine-grained layers are always thinner and rarely exceed 5 to 20 cm. In the garnet kinzigites of Masku the coarse-grained layers are about 30 cm. thick and contain abundant garnet, which is from 1 to 1.5 cm. in diameter. The fine-grained layers are of the same thickness as in the normal kinzigites, but some layers are thicker—*e.g.*, south of Askainen and northwest of Masku church. Figure 23 shows a kinzigite common in the parish of Rusko. The sandy fine-grained layers 2 to 10 cm. thick alternate with layers 5 to 50 cm. thick and rich in cordierite and garnet.

The layering of the volcanic rocks south of Vehmalainen is described above. Similar structures can be recognized with careful observation also in the areas of more intense metamorphism and deformation. The alternation between finer-and coarse-grained layers or lighter and darker ones are common. The uralite and plagioclase porphyrites may preserve their porphyritic structure up to the stage of intense migmatization. The phenocrysts are always strongly granulated.

Layering due to intercalation becomes very clear in the contact zones of the volcanic rocks and kinzigites. Thin leptite layers occur also elsewhere in the kinzigites—*e.g.*, 1.5 km northeast of Rusko church several layers of amphibole-bearing

leptite 5 to 10 m. thick are intercalated with kinzigite. They strike N. 60°–70° W. About 1.5 km. east along the strike thicker leptite layers occur. A volcanic complex with leptites and plagioclase porphyrites ½ km south of this zone suggests that these layers might be tuffs which originated at the same source as the volcanics and are intercalated with the sediments.

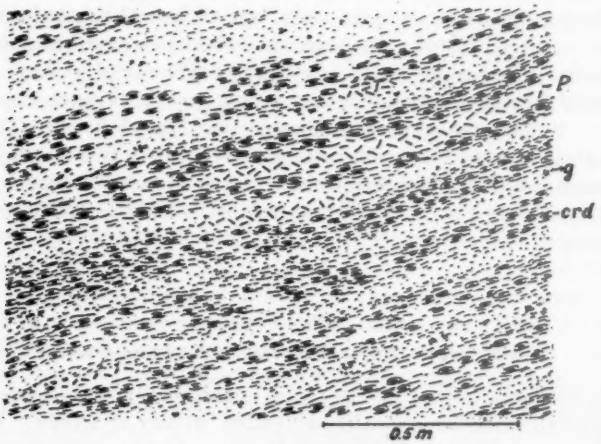


FIGURE 23.—*Bedding in the kinzigite in Rusko*
Cordierite- and garnet-bearing layers (crd, g) alternate. West of no. 1132, Pl. 1.

Large folds are shown on the map. Contacts between the folded older rocks and the older intrusives parallel the bedding. The folding is therefore also illustrated by the trend of contacts. The large anticlines northeast of Mynämäki and in Rymättylä as well as the synclines southeast of Masku are conspicuous. Strike and dip of bedding and planar structures in the vicinity of the contacts confirm their direction. It is thus possible to draw correct contacts where the contact itself is not exposed.

Small folds can be seen where the bedding is well preserved. As a rule the apices of the larger anticlines and synclines show many small folds. Their axes parallel those of the larger folds. There are, however, small folds whose axes differ from the general trend and are usually steeper. These occur in areas of steep dips on the flanks of the larger folds. They are considered to be secondary and were most likely formed during migmatization. (*See also* Hietanen, 1943.) The axes of the large folds strike about N. 60° E. The dip varies from 60° E. to 25° W., and the eastward dips of about 30° are prominent.

Lineation becomes visible in the parallel orientation of biotite flakes and hornblende needles. As a rule it parallels the axes of the main folds. However, where secondary folds occur lineation is steep and coincides with the direction of secondary axes. Lineation has been measured and plotted on the map, and in view of the described parallelism it indicates the direction of the prominent fold axes.

Cleavage usually parallels bedding, but in the apex of the folds it often transects the beds, remaining parallel to the general trend. In areas of intense migmatization cleavage is the only distinct structure, and bedding and lineation are either weak or lacking. In these localities where only banding, due to migmatization, is visible it is plotted with the cleavage symbols, but since bedding and cleavage are parallel it may indicate also the position of bedding. This is obviously true where cleavage is folded similarly to the bedding in adjacent areas. For instance, in a peninsula north of Naantali the cleavage plane curves from N. 60° E. through N. to N. 60° W. and further to N. 70°E. Bedding was observed only in a few outcrops here, and in each case the cleavage parallels it. It seems therefore safe to assume that this *s*-plane is really the original bedding. Similarly, in the parish of Mietoinen, only one *s*-plane occurs, and in most cases it probably parallels the bedding even if the intense migmatization has obliterated the primary structure. The occurrence of the long narrow amphibolite lenses and intrusions of the rocks of the trondhjemite series supports this view.

STRUCTURE OF THE SYNTECTONIC INTRUSIVES OF THE FIRST CYCLE

Foliation and the linear parallelism are common in the intrusives. Basic clots may occur in addition in the intermediate and acidic rocks.

In the long, narrow, probably sheet-shaped intrusives the foliation always parallels the contacts. Since the intrusives have penetrated the older rocks parallel to the bedding planes this foliation also parallels the bedding of the older rocks. This parallelism is clear in the contact zones of larger intrusive bodies, but in their central part only linear structures can be observed, or the rock is structureless.

In the long narrow lenses and in the contact zones of the larger intrusives linear structures parallel the fold axes of the older rocks. In the center of the Merijärvi charnockite pluton the linear parallelism is vertical or dips steeply toward the center. The foliation in the border zones dips steeply inward. In the eastern part of the pluton it is vertical, and it strikes from N. 70° E. through N. to N. 45° W. and W. Thus the pluton seems to be a cone. However, its western part is strongly elongated in the direction N. 70° E. Especially the southwestern part shows such a shape, and the foliation is distinct and parallels the contacts. This part consists of coarse-grained diorite-trondhjemite and ends in a long narrow lens folded with the kinzigites. In the center of the pluton south of Merimasku church the lineation dips steeply northwest like the foliation in the southeastern contact zone. These structures are thought to be partly the result of igneous flow. All planar structures in the intrusives, and wall rock, parallel the contacts, and the author therefore concludes that they are thus partly due to the flowage and partly to regional deformation. Intrusion took place in a late phase of folding. Regional deformation was not strong enough to destroy the flow structure of the igneous body in its center. After solidification the charnockite was more competent that the surrounding kinzigite, which yielded to the deformation more readily. The plutons in Luonnonmaa and those east of the Masku church show foliation planes which are folded in a way similar to bedding in the kinzigites. Each of these plutons occurs in the apex of a fold.

Dark inclusions are common in the charnockites and in the related trondhjemites.

They are either elongated or angular and consist of more basic rocks or accumulations of ferromagnesian minerals. Additional inclusions of older rocks may occur in the contact zones. Basic clots are common in the charnockites of the northernmost part of Luonnonmaa, in a road cut near Miekkilä in southern Luonnonmaa (no. 1111), in the road cut between Rusko and Killainen, about ½ km from the fork of the road, and

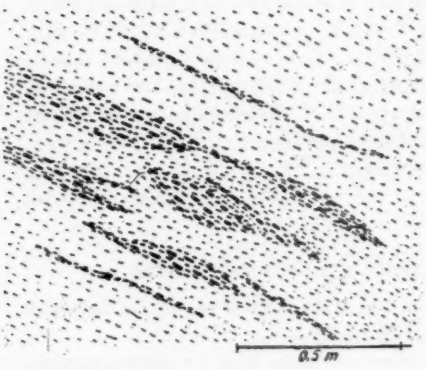


FIGURE 24.—*Inclusions of volcanics in granite-trondhjemite near Lake Hirvijärvi*

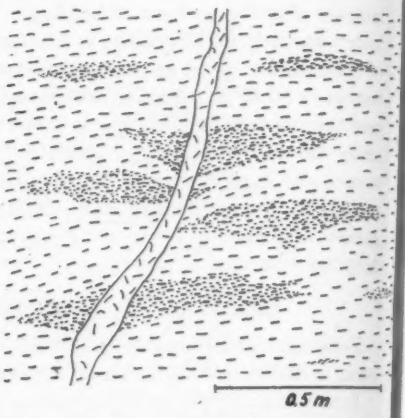


FIGURE 25.—*Younger pegmatite vein transects the granite-trondhjemite and the dark inclusions*

in several other localities near the contacts. Most of the clots are rounded or oval and range from 2 to 100 cm. in diameter. In places of strong foliation or lineation they are elongated parallel to these structures. The minerals in them are usually the same as in the main rock, but the amount of the light minerals is much less. The light-gray trondhjemite in central Luonnonmaa contains rounded and angular inclusions, composed mainly of biotite. These dark inclusions are most likely clots of the more basic rocks, and the minerals are altered according to Bowen's reaction principle.

Inclusions of the older rocks are mainly in the contact zones but may also occur in the center, as for instance in the granite-trondhjemites around the lake Hirvijärvi (Figs. 24, 25). Here the long narrow inclusions in the granite-trondhjemite parallel the foliation and represent various layers of older volcanic rocks. About 3 km south of the Masku church charnockite contains schlieren-shaped inclusions of kinzigite. These grade into the charnockite. The centers are rich in biotite and contain garnets. The amount of biotite decreases toward the charnockite, which contains in addition to these inclusions also schlieren which are only slightly richer in biotite than the bulk of the rock.

STRUCTURE OF THE INTRUSIVES OF THE SECOND CYCLE

The structures of the red microcline granite and associated rocks differ greatly from those of the older intrusives. They do not show the general conformity of planar and linear structures with the wall rocks that is characteristic of the trondhjemites and

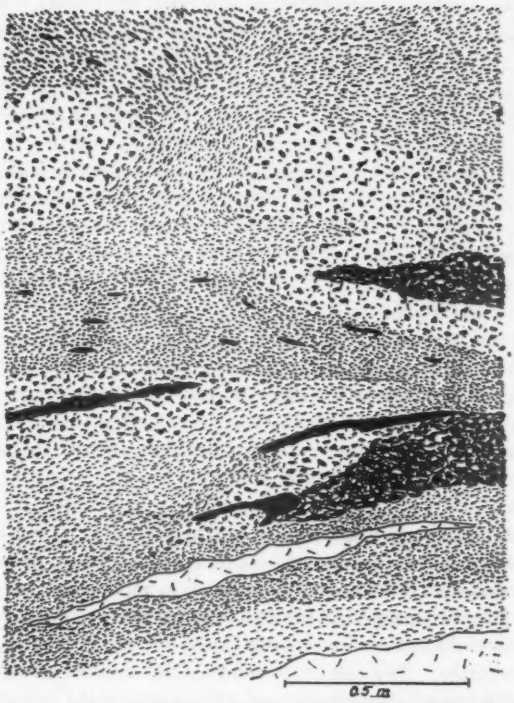


FIGURE 26.—*The rocks of the microcline granite series penetrating leptites*
Leptite has yielded to the force of the expanding magma, northeast of no. 1235, Pl. 1.

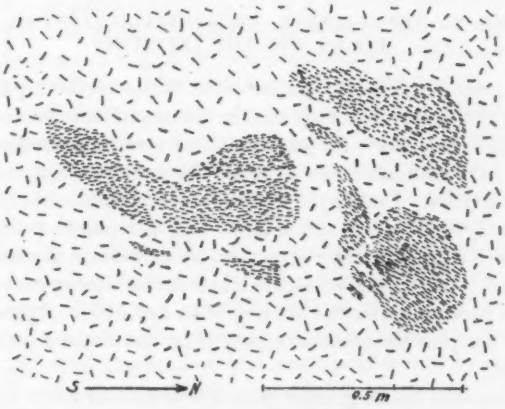


FIGURE 27.—*Inclusions of amphibolite in microcline-granite*
Near the farmhouse Häppilä in western Hirvensalo (west of no. 1334, Pl. 1).

charnockites, but are mostly structureless and often discordant. They may occur as long thin sheets parallel to bedding or cleavage in the older rocks (Pl. 1). This does not prove a syntectonic intrusion, but rather an invasion of the earlier-deformed rocks parallel to existing planes of weakness similar to the arteritic veins in the migmatites. (*Cf.* Hietanen, 1943.) Many of the small sheets of red microcline granite show an internal planar structure parallel to the walls. This sturcture becomes visible in the parallel arrangement of the (010) faces of feldspar phenocrysts twinned according to the Carlsbad law. As a rule the parallel orientation is common only in the narrow lenses and near the contact zones of the larger bodies, while their centers are oriented at random. This sort of parallelism of the planar structure and walls of an intrusion is characteristic of magma flows (Balk, 1937). The feldspar twins must have crystallized before or during the emplacement of the granite, while the other minerals, which show a random orientation, crystallized later.

The red microcline granite cuts the basic members of the same series and is thus clearly younger. The contacts between the basic members and the older rocks are conformable so that the foliation of the kinzigites and leptites parallels the contacts. A sketch of a small basic intrusion in the leptite 9 km east of Rusko (Fig. 26) shows this very well. The volcanic leptite was flexible and yielded to the force of the expanding intrusive material. The same can be seen on a larger scale in the contact zones of the granodiorite of Pansio north of Turku (Pl. 1). Here the older rock is kinzigite, and its schistosity parallels the contact. Minerals in the granodiorite show a random orientation. The hornblendite and gabbro north of Jäkärlä show a marginal platy structure parallel to the contacts, but their centers are structureless.

Angular wall-rock inclusions are common in the microcline granite, *e.g.* in the western part of Hirvensalo. The older rock is here a fine-grained schistose amphibolite. Dikes and small masses of medium-grained red granite cut it north of the actual contact. Angular and rounded inclusions of the same amphibolite are numerous in the granite along the contact zone (Fig. 27).

ROCKS DERIVED BY THE INTENSE RECRYSTALLIZATION OF THE EFFUSIVE ROCKS AND THE LEPTITES

The mineral assemblage of the effusive rocks of the northern part of the present area shows that they were metamorphosed in the amphibolite facies. The texture is wholly granoblastic, even where the porphyroblastic structure is megascopically visible. In many cases recrystallization has increased the grain size and obliterated most of the primary structures. Uralite porphyrite can thus alter into gabbros or diorites, plagioclase porphyrites into trondhjemites, and the potash-bearing leptites into granites. In each case some relict structure, which betrays the origin of the rocks, can still be found. Large plagioclases, in which earlier zoning is visible as inclusions in certain zones, occur in the diorite like nos. 988a and 1000. Aggregates of chlorite and serpentine are pseudomorphs after augite in the biotite-bearing diorite-trondhjemite no. 1010.

Leptite changes into coarse-grained granodiorite in a lens west of no. 1013. The southern part looks like igneous rock, while the northern part consists of less-altered leptite. It seems that some feldspar originated in the neighboring oligoclase granite

changing the leptite into a coarse-grained granodioritic rock. In this case the change would be related to the granitization, during which material was added, rather than only to recrystallization. The composition of the individual layers in the volcanic complex varies greatly, and it is often difficult to determine whether material has been added during metamorphism or even whether the rock is intrusive or metamorphic. This is true in some layers of the granite-trondhjemite north of Kalela. The southernmost and the central part of the granite consists of a coarse-grained well-foliated rock, whose structure megascopically resembles the coarse-grained sedimentary leptite. Another similar example is a small lens of trondhjemitic rock about 4 km east of no. 1027. Megascopically, and in mineral composition, it resembles fine-grained acidic layers in the leptite except that hornblende is lacking.

MIGMATIZATION

The rocks of the oldest complex show intense migmatization. The composition of the veinlets varies with the mineral composition of the host rocks and the neighboring intrusives. As in the Kalanti area some of the veinlets around the trondhjemite lenses consist of quartz and plagioclase and are thus trondhjemitic. Granitic veins are abundant in the oldest rocks but may occur in fewer numbers in some intrusives. As in the Kalanti area the veins parallel the best-developed *s*-plane in the rock.

The veins in the kinzigite-migmatites contain abundant cordierite and garnet; in each locality they are also present in the country rock. The essential minerals in these veins are quartz, microcline, and plagioclase. The occurrence of the cordierite and garnet and the absence of the minerals containing volatiles suggest the venitic genesis. Also in the volcanic rocks the vein material seems to originate from the country rock.

A good example of venitic migmatization of the volcanic leptites is seen in the area south of the village Vehmalainen and in Kalela in the parish of Karjala. As described in the section on Amphibolites and Leptites some leptite layers show a bedding. The beds have straight gradational boundaries as is common also in fine-grained sedimentary leptites (Pl. 2, fig. 5). In the areas of more intense migmatization this bedded leptite changes to a migmatite in which the veins consist of reddish medium-grained material and the older part of dark medium-grained rock. The minerals in the veins are the same as those in the light bands of the fine-grained leptite, and the older part consists of the same material as the dark layers. The boundaries of the veins are irregular as is usual in migmatites. Doubtlessly the vein material has been derived from the rock itself. During recrystallization metamorphic differentiation took place; feldspar and quartz also in the dark bands became mobile and were added to the similar material of the light bands. Locally this salic material forms broader veins and small masses, in which there are only schlieren of the dark material. Also in this case the origin of the rock can still be identified: part of the plagioclase and hornblende occurs as larger partly idiomorphic grains similar to the phenocrysts in the volcanic rocks. The formation of the larger pegmatites, which are common in the kinzigites and which contain abundant cordierite and garnet, can be explained in a similar manner. They also are venites. Many local granitizations of the older rocks may be correlated with this.

GRANITIZATION

The term granitization is here used for those processes in which quartz and feldspars or in some cases only potash feldspar are introduced into the rock. The rocks of the volcanic complex have changed into granites or diorites depending on the amount of added acidic material. The kinzigites alter into cordierite and garnet-bearing granites, the trondhjemites into granite-trondhjemites or granites. The process in which the volcanic amphibolites were changed into diorites could also be called a "partial granitization."

Granitization of the hornblende-bearing rocks of the volcanic origin can be studied ½ and 4 km south of lake Hirvijärvi in the parish of Karjala (no. 958 and 1027, Pl. 1). The result of the granitization is a reddish-gray granite-trondhjemite composed of quartz, plagioclase, microcline, biotite, magnetite, sphene, and zircon. Microcline forms about 30 per cent of the total feldspar. Some of the quartz occurs as small rounded grains. The microcline forms large grains and often includes other minerals. Abundant lenticular dark inclusions parallel the cleavage and consist mostly of biotite. In addition there are larger inclusions in which the structure and minerals of the older rocks are preserved in the center but whose margins consist largely of biotite. Accordingly also the smaller inclusions are considered to be remnants of the older rocks. Another example of granitization of volcanics is about 1 km east of locality no. 1027. There the amphibole-bearing leptites have changed into oligoclase granites. The contact zones in all these areas are gradational so that the amount of the salic minerals increases toward the granites. First only scanty separate feldspar grains appear in the dark rock rich in amphibole. Gradually their number increases, and the grains grow larger. Hornblende and biotite occur in this transition zone in newly crystallized grains of medium size. From the granite toward the amphibolite the percentage of plagioclase decreases from about 40 per cent to 30 per cent, and further to 22 per cent. At the same time hornblende alters to biotite. In the transition zones biotite often megascopically still shows the external habit of the hornblende. Thus the basic fine-grained leptites gradually change first to diorites and granodiorites and then to granites or granite-trondhjemites. The reaction hornblende → biotite requires potash, and accordingly the amount of the potash feldspar in these rocks is less than in the normal granites. However, the amount of microcline seems to depend in the first place on the amount of this mineral in the magma that has caused the granitization. This becomes clear in the area 3 km east of the Rusko church. Here the red microcline granite has caused granitization of the amphibole-bearing leptite, and the bulk of the feldspar in the new rock is perthitic microcline similar to that in the red granite, the pegmatitic portion of which occurs in the same area. Also this pegmatite contains hornblende originating from leptite. It is visible as green patches in the coarse-grained red rock and is associated with chlorite and magnetite.

The dark-gray medium-grained gabbros and diorites and lighter-gray diorite-trondhjemites in Paattinen and north and east of it may represent different stages of granitization of volcanic amphibolites as well as intrusives. In many cases it is even clear that the older rocks—the amphibolites—form the main part of the rock and that only minor amounts of quartz and feldspar were added during the recrystallization. Usually the diorites of this type are finer-grained and more distinctly schistose than

the intrusive diorites. In addition they contain unaltered layers, inclusions, and schlieren of the older rocks. This is clearly visible 4 km north of the village Karviainen, Aura (SW corner, Fig. 2) where the coarse-grained light-colored trondhjemite similar to the "drop-quartz" type of the Kalanti area (Hietanen, 1943) penetrates the volcanics. Every transition from this light rock to medium-grained diorite trondhjemite and diorite and from diorite to fine-grained amphibolite and leptite is represented. The volcanic amphibolites and leptites form the main part of the older complex in this district; the kinzigites are subordinate. Wide areas are occupied by diorite, and many of them seem to be due to partial granitization of volcanics. However, the presence of a dioritic magma must also be assumed, because diorites invade the kinzigites as well as the volcanics. Another question is how this magma was formed. Also it may well contain large amounts of volcanic material. The three following factors must be considered in the formation of the "intrusives" of the present district: the presence of a magma, palingenetic processes, and granitization. Granitization may be partial or complete. Palingenesis and granitization are common processes in the areas where erosion has been deep, and both have been widely discussed in the literature.

As pointed out, the products formed by the granitization processes vary according to the primary rock and the degree of granitization. In addition the active magma may have been granitic or trondhjemitic. The introduction of granite with abundant potash feldspar was later than that of granitic material poor in potash. Thus diorites formed by partial granitization may have undergone another granitization during which potash feldspar was added. This later granitization affected the rocks only locally, and the potash feldspar grains occur in cloudlike swarms, whereas the earlier partial granitization affected wider zones and lenses parallel to the *s*-planes. It appears thus more like a contact phenomenon, or a phenomenon connected with strong differential movements in certain zones. This earlier granitization took place during the orogenic period, contemporaneous with the intrusion of the magma from which the trondhjemites, the charnockites, the Pöytyä diorite, and the associated intrusives were derived. The potash-rich solutions came much later after the main period of movements ceased, obviously contemporaneous with the younger intrusives, which as a rule are rich in potash, and these solutions invaded the older formations metasomatically.

The granitization products of the kinzigites are described in the section on Kakola granite.

The granitization of the trondhjemites is common in the northern and eastern part of the area. Only potash feldspar seems to be added in the rock in this case, which thus is analogous to the *Kalifeldspatisierung* in the Kalanti area (Hietanen, 1943, p. 96). The occurrence of the younger microcline as thin films between the other minerals and as antiperthitic spots in the plagioclase suggests migration by means of a fluid phase between the grains as well as diffusion through the crystal lattices.

KAKOLA GRANITE

The granites which contain patches of cordierite and garnet regularly distributed throughout the rock are called Kakola granites (Pl. 4, figs. 5, 6) after their type lo-

cality in the Kakola hill in Turku. Like the kinzigites, Kakola granites may contain only cordierite or only garnet or both. The distribution of these minerals is very irregular so that the same outcrop may consist partly of cordierite granite and partly of garnet granite. Also the grain size and the color vary greatly. Most of the cordierite Kakola granites are reddish, coarse-grained, rich in microcline, and poor in biotite. In the pegmatitic varieties the biotite is absent. The garnet and the garnet-cordierite granites are reddish or gray, usually medium-grained, and contain biotite, similar to the normal granites. Most of the Kakola granites in the city of Turku and many occurrences around it belong to this group. They, like the normal granites, show good parting planes and are therefore quarried for building stones. Zircon, apatite, and some monazite are common accessories in all Kakola granites. Laitakari (1934) reports also a few andalusite and dumortierite grains.

The genesis of these peculiar granites is clearly connected with the granitization. Their mineral assemblage is derived partly from the kinzigites in which they occur, and partly from the granitic material which caused the granitization. The amount of either depends on the local circumstances. In some cases Kakola granite may contain only kinzigite material. This type is seen in the railroad cut 2 km southwest from Raisio church. Here the kinzigite is an intensely migmatized garnet rock. The veins are 1 to 20 cm. thick and consist of medium-grained, light-gray granite containing abundant garnet. Garnets are rounded or partly idiomorphic, 1 to 2 cm. in diameter, and evenly distributed. Locally these veins widen, and small masses of a similar garnet-bearing aplitic granite occur in the kinzigite. This granite is most likely venitic and formed by metamorphic differentiation according to Eskola's solution principle (Barth, Correns, and Eskola, 1939).

In this locality the masses derived by metamorphic differentiation from the kinzigites are rather small, and their composition is near the pegmatitic material common also in the veins. Biotite is absent in both. True Kakola granites contain biotite. Their formation must therefore be somewhat different from the venites described above. The outcrops in a road cut through the hill Samppalinnanmäki in southern Turku throw further light on this problem. The main part of the rock consists of a light-gray, fine-grained biotite granite. It contains large inclusions and schlieren of another granite richer in biotite. There are also many small masses and veins of medium-grained reddish granite, which contain patches of garnet but no bioitite. It seems to the writer that the fine-grained biotite-bearing parts were derived from the fine-grained beds of kinzigite by recrystallization and partial granitization. Granitization was caused by the red microcline-rich granite. The red parts contain more granite and less kinzigite material. The increase of the granite-pegmatitic component is visible in the field in the lighter and more reddish color of the granite.

As a rule the larger Kakola granite bodies appear more like normal granites with biotite evenly distributed throughout. Their composition would be approximately the same as the average composition at Samppalinnanmäki. Their granitic structure indicates that the rock was recrystallized and all the older structures were obliterated. These rather homogeneous granites show how granitic textures are attained as soon as the composition of the material derived by the metamorphic and palingenetic processes becomes granitic. The smaller bodies and veins are usually coarse-grained

and close to pegmatites in composition. Coarse-grained parts containing only cordierite are locally found also in larger bodies. In some outcrops, as in eastern Turku, cordierite is changed into chlorite, and biotite surrounds these altered cordie-

TABLE 6.—*Chemical and mineralogical composition of a red (no. 1) and gray (no. 2) Kakola granite from Turku*

	1	2		1	2
SiO_2	73.56	73.50	quartz	30.48	33.51
TiO_2	0.11	0.19	plagioclase	An_{14} 29.13	An_{22} 29.31
Al_2O_3	14.72	13.59	microcline	33.20	30.20
Fe_2O_3	0.39	0.57	biotite		1.84
FeO	0.75	1.27	muscovite	1.90	
MnO	trace	0.04	garnet	2.00	3.40
MgO	0.20	0.40	cordierite	1.26	
CaO	0.83	1.42	sillimanite	0.83	0.23
Na_2O	2.99	2.68	ilmenite	0.16	0.26
K_2O	5.78	5.25	hematite	0.26	0.24
P_2O_5	0.03	0.00	monazite	0.13	
H_2O+	0.56	0.29	$+H_2O$	0.83	0.44
H_2O-	0.28	0.23			
	100.20	99.43		100.18	99.43

	NORMS	
	1	2
qu	31.38	33.46
ab	25.40	22.71
an	4.11	7.05
or	34.28	31.00
cor	1.99	0.90
en	0.50	0.99
hy	0.87	1.62
mon.	0.09	
il	0.21	0.36
mt	0.56	0.83
H_2O	0.84	0.52
	100.23	99.44

	NIGGLI VALUES	
	1	2
si	422.5	416.5
ti	0.48	0.76
al	49.90	45.50
fm	7.00	12.10
c	5.13	8.65
alk	37.97	33.85
qz	170.6	181.1
k	0.56	0.56
mg	0.25	0.28
c/fm	0.73	0.71

rite patches as a reaction rim. The absence of cordierite in many garnet Kakola granites north of Turku may be partly due to this reaction, which would point to a lower temperature and the presence of water during the granitization.

For a quantitative picture of the composition and the amount of granitic material added during granitization, two Kakola granites were analyzed chemically by Mr. O. von Knorring (Table 6). Analysis no. 1 was made from a red coarse-grained cordierite-garnet granite, and no. 2 from a gray, medium-grained type containing biotite and garnet. The analyses are very similar. The red Kakola granite contains

a little more potash and alumina due to abundance of potash feldspar and presence of cordierite and sillimanite. If the Kakola granite is a granitized kinzigite we can divide the analyses in Table 6 into two parts to find the amount and composition of the added material. The composition of the kinzigite used for this purpose is de-

TABLE 7.—*Amounts of kinzigite and granitic material in Kakola granites*

	1 red Kakola granite Turku		2 gray Kakola granite Turku (1138)	
	Recalculated composition	Granitic material	Recalculated composition	Granitic material
kinzigite	12.52		22.74	
quartz	27.79	31.76	26.88	34.80
plagioclase	An_{13} 25.13	An_{13} 28.73	An_{24} 23.00	An_{24} 29.75
microcline	29.77	34.03	27.30	35.35
muscovite	4.43	5.06		
sillimanite	0.03			
monazite	0.09			
hematite	0.10		0.03	
rutile			0.05	
water	0.11			
	100.00	100.00	100.00	100.00

termined as an averge composition of the kinzigites from Kalanti (Hietanen, 1943, p. 16, analyses 1–6). The results of the calculation are given in Table 7.

In the gray Kakola granite there is about 23 per cent kinzigite and 77 per cent granite-pegmatitic material. In the red variety the amount of kinzigitic material is only 12.5 per cent, and the granite-pegmatitic part, which is somewhat richer in both alkalies than in the gray variety, is 87.5 per cent of the total. In both the percentages of quartz, plagioclase, and microcline in the granite-pegmatitic material are about the same. In the red Kakola granite the plagioclase is more albitic. Its composition corresponds well with that of the plagioclase in the red microcline granite, while the plagioclase in the gray biotite-garnet-bearing Kakola granite has approximately the same composition as that in the kinzigites. This supports the view that the gray variety contains more venitic material and the red type more solutions from microcline granites.

Field examination of the Kakola granites leads to the same conclusion. As a rule the end product of granitization depends mainly on the character of the magma which causes it. Accordingly there should be several types of Kakola granites associated with the various granites and various kinzigites. On the road Kuninkaantie 1 km north of Raisio a red microcline granite contains patches of garnet and cordierite. This type occurs also east of Turku with a gray variety. It seems that the main part of the material in this rock came from the microcline granite magma. Most Kakola granites north of Turku are gray and medium-grained, while the light-reddish somewhat coarser varieties occur with the gray ones in other localities. If we bear in mind the rock of Samppalinnanmäki the difference between these two types could be

explained as follows: the gray, finer-grained types contain more kinzigitic material, while the coarse-grained reddish varieties have taken up more granitic material. In many cases, however, the red, coarse-grained cordierite-garnet granites have the same composition as that material which consists of most soluble substances containing in addition cordierite and garnet. These venitic Kakola granites are always pegmatitic.

The garnet granites associated with the charnockites in Lieto and in the Satava zone clearly contain, in addition to remnants of the kinzigites, granitic material of two ages. The center of the garnet-bearing granite area in Lieto shows in many places features characteristic of the granite-trondhjemites belonging to the charnockite series. The amount of garnet and biotite increases greatly toward the border, and the color changes from brownish to bluish-gray resembling that of the kinzigites. The pegmatites are rich in potash feldspar and have invaded the rock in veins and masses.

In Satava zone and in the small islands south of Rymättylä this two fold granitization is perhaps still more obvious. In many localities the grayish garnet-bearing granite is penetrated by numerous red pegmatitic veins which also contain garnets. The older grayish granite grades into the red granite, and most of the second granitization has gone so far that the rock is now completely homogeneous. In such cases the older granite may have contained potash feldspar, and its color is not very different from that of reddish pegmatites of the second cycle. There are, however, also examples where the older acidic rock consisted of bluish-gray trondhjemite. On the northern side of the grave yard about 2 km south of Turku the red Kakola granite includes numerous rounded lenses of fine-grained gray rock. Their size and their color vary. Some are bluish-gray and consist of a trondhjemite which looks like the fine-grained gray trondhjemite about 1 km west of the same locality. Some others are reddish-gray, and microscopic examination shows that they contain also potash feldspar. These reddish-gray inclusions (primarily trondhjemitic) were partly granitized during the invasion of the red grantitic material.

If is often difficult to tell what role each of the three possible processes—(1) extraction and redeposition of the most soluble substances, (2) addition of pegmatitic material, and (3) palingenesis—have played during the formation of these rocks. In addition the granitic material which was added may have come during the intrusions of first or second cycle. It apparently depends on the local circumstance which process has been most important. The addition of abundant pegmatitic solutions of red microcline granites would cause the formation of red Kakola granites. The gray varieties probably formed partly by metasomatic addition of pegmatitic material and partly by metamorphic differentiation and palingenesis. The light-reddish pegmatitic types may be venites, or, in the other localities, they may have derived most of their material from the granite-trondhjemitic or granitic magma. To this type belong the long narrow lenses in the northernn part of the area. Many granite-trondhjemites in the kinzigites contain cordierite and garnet and sometimes also sillimanite (locality 920) and are thus analogous to the Kakola granites. The trondhjemites may locally contain these minerals, especially in their contact zones. Most Kakola granites were obviously formed by metamorphic and metasomatic processes

during which feldspars and quartz were added, and thus they never existed as a magma. On the other hand the existence of a Kakola granite magma must be assumed, because Kakola granites occur also in areas where no cordierite or garnet exist—*e.g.*, they penetrate the amphibolites and leptites in many localities. At least a part of the Kakola granite magma thus became mobile and intruded the neighboring formations as veins and small masses. The processes which caused the formation of the Kakola granite magma would then be as follows: venites containing cordierite and garnet were formed by selective resolution and redeposition during the high-temperature metamorphism and differential movements. Pegmatitic solutions from outside were added during the intrusions of acidic rocks. Parts of the material which attained a granitic composition during these processes became mobile.

The garnet granites may be petrographically compared with the garnet-bearing gneisses of Middle Sweden. These gneisses are considered intrusive by Magnusson (1936), and the garnet crystallized because of the enrichment of iron in the acidic rest magmas. A similar explanation can be used for the occurrence of abundant garnet in the plagioclase pegmatites which cut the trondhjemites in the Merimasku-Naantali-Turku zone. The trondhjemites contain only a few small garnets and occasionally some cordierite, but the pegmatitic portions, which are clearly formed from the rest solutions of the trondhjemitic magma, contain locally abundant idiomorphic garnets 1 to 2 cm. in diameter. However, also in this case it seems clear that the garnet originally came from kinzigites, which were assimilated by the trondhjemitic magma.

SUMMARY AND CONCLUSIONS

The rocks of the Turku district can be divided into four main groups according to their age: (1) the oldest complex, (2) the intrusives of the first cycle, (3) the intrusives of the second cycle, (4) post-tectonic intrusives.

1. The oldest complex consists of intercalated gneissic rocks of sedimentary and volcanic origin, folded together and migmatized. The rocks of sedimentary origin usually contain cordierite and garnet and are called kinzigites. Their bedding can still be observed at many localities in the alternation of coarse- and fine-grained layers and in the distribution of cordierite and garnet. Later migmatization, however, has mostly obliterated primary structures, and only the banding, due to veining which may parallel the original bedding or later foliation, remains visible in the field.

In only a few localities are primary structures of the volcanics preserved sufficiently to provide clues to their origin. Relict structures and their comparison with exposures in which primary structures are well preserved prove that the hornblende-bearing rocks north of Turku consist almost exclusively of metamorphosed lavas and tuffs, while the amphibolites south of Turku are most likely sedimentary.

The alternate layering of tuffs, lavas, and sediments suggests several successive volcanic eruptions, and the thickness of the sedimentary layers provides data on the length of time between eruptions.

The base of the sedimentary and volcanic complex has not been found; there are not even any conglomerates which might indicate proximity of basal formations. The sedimentary leptites within the volcanic complex of Kalela contain small pebbles of

somewhat bluish quartz which looks like the quartz in the gneissic, rather inhomogeneous trondhjemitic rocks in the anticlinal core of the same area. However, the structures of these rocks are similar to the older intrusives, and their concordant contacts render more definite determinations impossible. It may well be that these oldest intrusives contain material from pre-kinzigite and pre-volcanic formations which became granitized and newly mobilized during the orogeny.

2. The intrusives of the first cycle are divided into three main groups according to their mineral assemblage: (1) the trondhjemite series, (2) the charnockite series, and (3) the Pöytyä diorite and associated rocks. Each of these comprises a complete differentiation series which grades into the others. The chemical compositions of corresponding members of each group are analogous. Many charnockites and trondhjemites show stuctures of igneous rocks, but some occurrences are clearly formed metasomatically or metamorphically. The juvenile material which caused the formation of charnockitic and trondhjemitic rocks originated in the same source. Thus the reactions between the older rocks and the invading magma or solutions caused the differentiation of the intrusives of the first cycle. Absence or presence of water, the temperature, and the pressure seem to be the determining factors during the differentiation of the charnockitic parent magma.

The area of the pyroxene gneisses south of Turku may well represent the cradle of the charnockitic magma. Intrusions in the immediate vicinity remain charnockitic, but away from the original magma chamber watery sediments and volcanics were assimilated, and the differentiate became trondhjemitic. The dioritic rocks in the parish of Pöytyä were increased in volume by the partial granitization of the basic volcanics which there form the main part of the older rocks.

3. The intrusives of the second cycle differ chemically and structurally from those of the first cycle. The intrusive cycle includes all members of a complete differentiation series from hornblendites and gabbros to granites and pegmatites, but granites prevail. The acidic members are rich in microcline and transect the older rocks discordantly.

Considering the magmatic activity of the entire orogenic period (first and second cycle) we see that the magmas show a common tendency: the younger magmas are enriched in potash and also in iron. This feature is common also in other areas in southwestern Finland. The older intrusive series are, as a rule, rich in calcium and magnesium and poor in potash, whereas the younger granites are rich in potash.

4. The emplacement of the diabase dikes, the anorthosite of Kolinummi, and the rapakivis is later than the intrusives of the second cycle and the general migmatization and granitization.

The present area offers suitable material for the study of the formation of trondhjemitic and granitic magmas. Metasomatism, anatexis, and palingenesis have played an important role. Primary magmas furnished solutions and heat during regional metamorphism. The solutions invaded the older rocks causing their migmatization and granitization.

The migmatization of the oldest complex started during the first cycle and continued during the second cycle, and in many localities the rocks have undergone a twofold migmatization. The migmatite veins were derived partly by metamorphic

differentiation and partly by addition of juvenile material. Since the juvenile material may be trondhjemitic or granitic and since the oldest complex is rather inhomogeneous, the composition of the veins varies greatly. Differential movements preceded migmatization, but the intrusion of solutions continued beyond folding. Granitic textures originated in this later period.

Granitization is comparable with migmatization as to time, added material, and material derived from the older rocks. Whether the end product is trondhjemitic or granitic depends on the compositions of magma and host rocks.

The granitization products of the kinzigites are called Kakola granites. Like the migmatization their formation started during the first and continued into the second intrusion cycle. (1) Extraction and redeposition of the most soluble substances, (2) addition of the pegmatitic material, and (3) palingenesis seem to have a part in their formation. Addition of the solutions from microcline granites is suspected because of the red color. In the southern part of the area, where the influence of the red microcline granites is strongest, the Kakola granites formed during the first cycle were migmatized during the intrusion of microcline granites. Locally the end product is a homogeneous granite,and thus the rocks of the oldest complex have here undergone a twofolded granitization.

Granitization leads to homogeneous rocks, whereas during migmatization originally homogeneous rocks become very inhomogeneous; the light and dark minerals are accumulated in separate bands. There still is, however, a gradation from the highly metamorphic schists to migmatites, to migmatitic granites (or trondhjemites), and finally to granites. This shows that the same processes must first lead to the formation of migmatitic rocks and later to the formation of granitic texture. During migmatization the pervading solutions penetrate the rock along planes of least resistance. At the same time a selective re-solution in the rock adds material to the veins. With rising temperature, the liquid portion grows, only a part of it is extracted into the veins, and the residual portion remains in the rock as "ichor" causing further changes metasomatically in still unaltered rock. Granitized inclusions of older rocks in the migmatic granites provide evidence of this stage. Finally the continued anatexis and introduction of solutions lead to the homogeneous rock. The composition of the end product is close to that of the active magma, but traces of older rocks occur. This must mean that the solutions which invaded the older rock metasomatically form an essential part of the end product. Thus its volume increases, and as a consequence the pressure rises, then the migmatic granite melts, and anally the country rocks are invaded by palingenetic granite. The Kakola granite veins cutting the amphibolites in many localities of the present area are a good example.

The charnockite magma may have formed similarly, and petrographic and field examinations make this most probable. In this case the pyroxene amphibolites may represent the older rocks, and the temperature must have been considerably higher than during granitization. Probably also in this case the sedimentary part forms an insignificant portion, at least in the charnockitic intrusions, but its influence becomes clearer in those parts of the magma which wandered farther from the magma chamber (trondhjemites, diorite of Pöytyä) and in all more acidic rocks of the series.

REFERENCES CITED

Adams, F. D. (1929) *The geology of Ceylon*, Can. Jour. Research, vol. 1, p. 425–511.

Asklund, B. (1925) *Petrological studies in the neighbourhood of Stavsjö at Kolmården*, Sveriges Geol. Und., ser. C, no. 325, p. 1–122.

Balk, Robert (1937) *Structural behavior of igneous rocks*, Geol. Soc. Am., Mem. 5.

Barth, T. F. W. (1936) *The crystallization process of basalt*, Am. Jour. Sci., 5th ser., vol. 31, p. 321–351.

———, Correns, C. W., and Eskola, Pentti (1939) *Die Entstehung der Gesteine*, Berlin.

Borgström, L. H. (1915) *Die chemische Zusammensetzung der Skapolite*, Zeitschr. Krystal., Bd. 54.

Bowen, N. L., and Schairer, J. F. (1935) *The system MgO-FeO-SiO₂*, Am. Jour. Sci, 5th ser., vol. 29, p. 151–217.

Buddington, A. F. (1939) *Adirondack igneous rocks and their metamorphism*, Geol. Soc. Am., Mem. 7.

Bugge, C. (1910) *The eastern part of Ellesmere Land and Prudhoe Land opposite across Smith Sound*, 2d Fram Exped., Rept., no. 22.

Bugge, Jens (1940) *Geological and petrographical investigations in the Arendal District*, Norsk Geologisk Tidsskrift, Bd. 20.

——— (1943) *Geological and petrographical investigations in the Kongsberg-Bamble formation* Norges Geol. Und., no. 160.

——— (1945) *The geological importance of diffusion in the solid state*, Norske Vid. Akad., Oslo. I Mat. Naturv. Kl., no. 13.

Eskola, Pentti (1914) *On the petrology of the Orijärvi region in southwestern Finland*, Comm. Geol. Finl., Bull. 40.

——— (1938) *On the esbiotic crystallization of orbicular rocks*, Jour. Geol., vol. 49, no. 3, pt. II.

Goldschmidt, V. M. (1920) *Geologisch-petrographische studien im Hochgebirge des Südlichen Norwegens. V Die Injektionsmetamorphose in Stavanger-Gebiete*, Kristiania Videnskr.-selsk. Skrift. I Mat.-naturvid. Kl., Bd. 2.

Groves, A. W. (1935) *The charnockite series of Uganda, British East Africa*, Jour. Geol., vol. XCI.

Hackman, Victor (1923) *Der Pyroxen-Granodiorite von Kakskerta*, Comm. Geol. Finlande, Bull. 61.

Hausen, H. (1944) *Geologische Beobachtungen im Schärenhof von Körpo-Nagu, Südwest-Finnland*, Medd. Åbo Akad., no. 24.

Hess, H. H. (1941) *Pyroxenes of common mafic magmas. Part I*, Am. Mineral., vol. 26, p. 515–535.

———, and Phillips, A. H. (1938) *Orthopyroxenes of the Bushveld type*, Am. Mineral., vol. 23, p. 450–456.

Hietanen, Anna (1943) *Über das Grundgebirge des Kalantigebietes im südwestlichen Finnland*, Acad. Scient. Fennicae, Ann., Serie A, III, no. 6; Comm. Geol. Finlande, Bull. 130.

Holland, T. H. (1900) *The charnockite series. A group of Archean hypersthenic rocks in Peninsular India*, Geol. Survey India, Mem. 28, pt. 2, p. 119–249.

Holmquist, P. J. (1921) *Typen und Nomenklatur der Adergesteine*, Geol. Fören. i Stockholm Förh., Bd. 43, Häfte 6–7.

Kolderup, C. K. (1896) *Die Labradorfelse des westlichen Norwegens, I. Das Labradorfelsgebiet bei Ekersund und Soggendal*, Berg. Mus. Aarbog, no. V.

——— (1903) *Die Labradorfelse des westlichen Norwegens. II. Die Labradorfelse und die mit denselben verwandten Gesteine in dem Bergensgebiete*, Berg. Mus. Aarbog, no. 12.

Lacroix, A. (1910) *Sur l'existence à la Côte D' Ivoire d' une série pétrographique combarable à celle de la charnockite*, Acad. Sci. Paris, C. R., vol. cl.

Laitakari, Aarne (1934) *Suomen kivien raskaista mineraaleista. III Kakolan graniitin raskaat mineraalit, with an English summary: Heavy minerals in Kakola granite*, Mineral. Geol. Inst. Teknillinen Korkeakoulu (Technological University) Helsinki, Finland, Papers, no. 16 (Tekn. aikakauslehti no. 12).

Lundegårdh, Per H. (1943) *The Grovstanäs Region. An Ultra-basic gabbro massif and its immediate vicinity*, Geol. Inst. Upsala, Bull. XXIX.

Magnusson, N. H. (1936) *Berggrunden inom Kantorps Malmtrakt*, Sveriges Geol. Unders., ser. C, no. 401.

Metzger, Adolf A. T. (1945) *Zur Geologie der Inseln Älö und Kyrklandet in Pargas-Parainen, S. W. Finnland*, Medd. Åbo Akad., no. 27.

Moberg, K. Ad. (1890) Explanation of geologic sheet No. 10, Geol. Comm. Finland.

Parras, Kauko (1941) *Das Gebiet der Pyroxen führenden Gesteine im westlichen Uusimaa in Südfinnland*, Geol. Rundschau, Bd. 32, Hefte 4–5.

Ramberg, Hans (1943) *En Undersøkelse av Vestrandens regionalmetamorfe bergarter*, Norsk geol. tidskr., Bd. 23.

———— (1945) *Petrological significance of subsolidus phase transitions in mixed crystals*, Norsk geol. tidskr., Bd. 24, Hefte 1–2.

Reinhard, M. (1931) *Universal Drehtischmethoden*, Basel.

Rogers, G. S. (1911) *The Cortlandt series near Peekskill, New York*, Am. N. Y. Ac. Sci. XXI.

Sander, Bruno (1911) *Über Zusammenhange Zwischen Teilbewegung und Gefüge in Gesteinen*, Tscherm. Mitt., vol. 30.

Sederholm, J. J. (1899) *Über eine archäische Sedimentformation im Südwestlichen Finnland*, Comm. Géol. Finlande, Bull. 6.

———— (1907) *Om granit och gneis*, Comm. Géol. Finlande, Bull. 23.

———— (1934) *On migmatites and associated pre-Cambrian rocks of southwestern Finland. Part III. The Åland Islands*, Comm. Géol. Finlande, Bull. 107.

Washington, Henry S. (1916) *The charnockite series of igneous rocks*. Am. Jour. Sci., 4th ser., vol. 41.

Wilkman, W. W. (1898) Explanation of geologic sheet No. 32, Geol. Comm. Finland.

Winchell, A. N. (1933) *Elements of optical mineralogy and introduction to microscopic petrography. Part II. Descriptions of minerals*, New York.

STANFORD UNIVERSITY, CALIF.

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, APRIL 9, 1946.

PLATE 2. —PHOTOMICROGRAPHS

FIGURE

1.—Derviation of biotite at the expense of cordierite. Cordierite (crd) shows an alteration to chlorite along the cracks and to biotite along the margins. Worm-shaped quartz grains are included in biotite. Cordierite-garnet kinzigite (no. 1115), southern shore of Luonnonmaa. (× 15)

2.—Biotite and quartz inclusions in microcline, and along margins in cordierite. The center of cordierite (crd) includes sillimanite. Microcline (mi) shows a few perthite lamellae. Cordierite kinzigite no. 892, Merimasku. (× 15)

3.—Plagioclase-uralite porphyrite. South of Vehmalainen (no. 926, Pl. 1; × 15).

4.—Ore amygdules in fine-grained leptite. The magnetite (black) is surrounded by small grains of sphene. Next layer consists of plagioclase and quartz. Femic minerals in the groundmass are hornblende and biotote. South of Vehamlainen (no. 929, Pl. 1; × 10).

5.—Layering in leptite. Light-colored layers consist of plagioclase and quartz. Biotite and some hornblende with these salic minerals form the darker layers. 300 m. north of no. 929, Pl. 1 (× 10)

6.—Cummingtonite-garnet amphibolite. Tammisluoto, west of Rymättylä (no. 1277, Pl. 1; × 15).

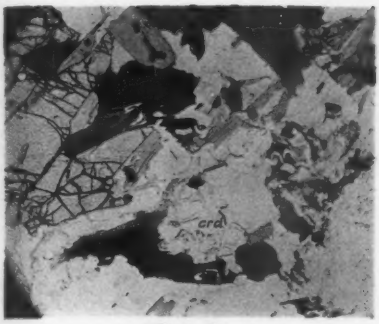


FIGURE 1

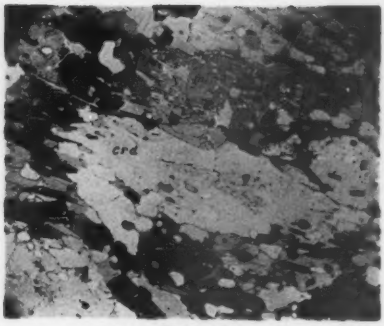


FIGURE 2

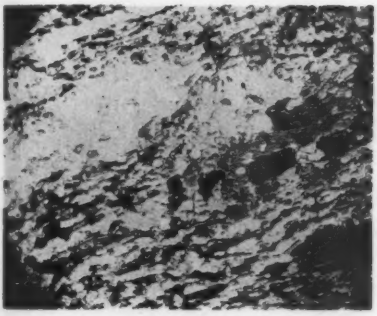


FIGURE 3

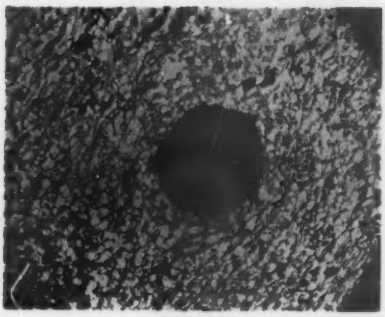


FIGURE 4

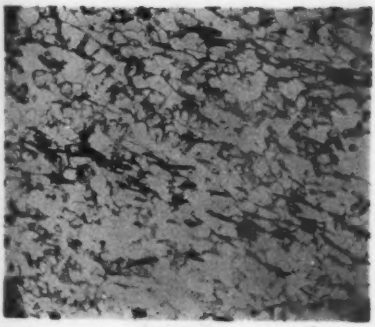


FIGURE 5

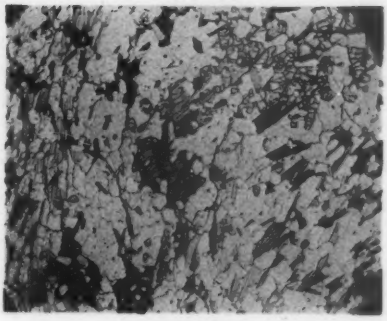


FIGURE 6

PHOTOMICROGRAPHS

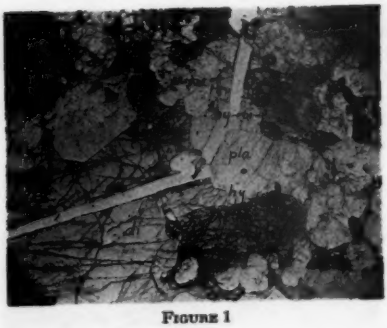


FIGURE 1

FIGURE 2

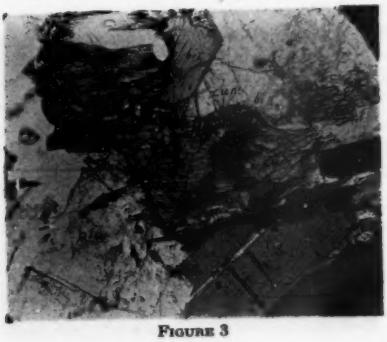


FIGURE 3

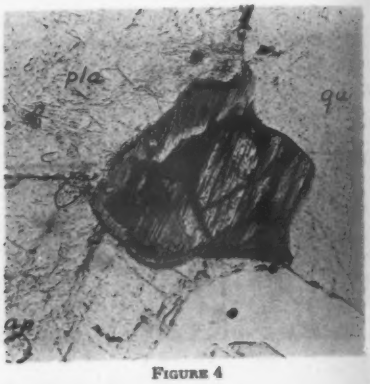


FIGURE 4

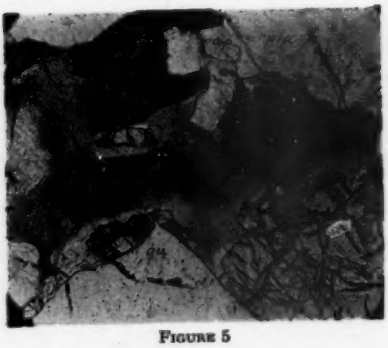


FIGURE 5

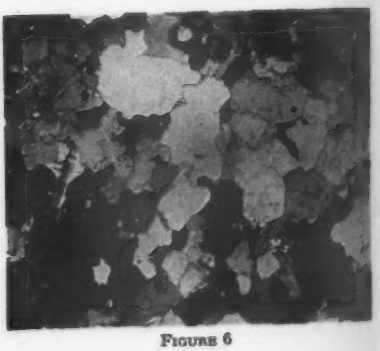


FIGURE 6

PHOTOMICROGRAPHS

PLATE 3.—PHOTOMICROGRAPHS

FIGURE

1.—Contemporaneous crystallization of pyroxenes in norite. Hypersthene (hy) and diopside (di) can be easily distinguished also without Nicol prism because of heavy cleavage and cracks in hypersthene. Lieranta, Merimasku (no. 877, Pl. 1; × 10).

2.—Alteration of hypersthene to biotite along cracks. From norite layer in Pahaluoto (no. 1294, Pl. 1; × 46).

3.—Complete destruction of hypersthene in acidic charnockite. Alteration products are biotite (bi II) and cummingtonite (cum). Diopside (di), which surrounds them, is unaltered. Biotite I is primary. West of Lenklahti, Rymättylä (no. 1087, Pl. 1; × 45).

4.—Alteration of hypersthene to hornblende (ho) and biotite (bi). Charnockite, 8 km northeast of Turku (no. 1240, Pl. 1; × 75).

5.—Hornblende partly surrounded by hypersthene. Charnockite, 4 km from Masku towards Turku. (× 46)

6.—Granite-trondhjemite. Note scarce microcline. Livonsaari (no. 905, Pl. 1; × 15).

PLATE 4.—PHOTOMICROGRAPHS

FIGURE

1.—Destruction of hypersthene to biotite (bi II) and hornblende (ho II). Granite-trondhjemite, Hintsholma, west of Kakskerta. (no. 1360, Pl. 1; × 17).

2.—Diopside lamellae in hypersthene parallel to (010). Anorthosite of Kolinummi, Yläne (no. 957, Pl. 1; × 16).

3.—Intergrowth of two hypersthenes. (041) face of the lamellae is parallel to (010) of the main mineral. Norite no. 1336 south of Turku (× 56).

4.—Quartz porphyry dike. Southern shore of Ruissalo west of Turku (no. 1356, Pl. 1; × 15).

5.—Accumulation of cordierite garnet in the center. Kakola granite, Turku (× 15).

6.—Kakola granite. The dark patches are accumulations of cordierite and garnet. Kako Turku. (× 1.18).

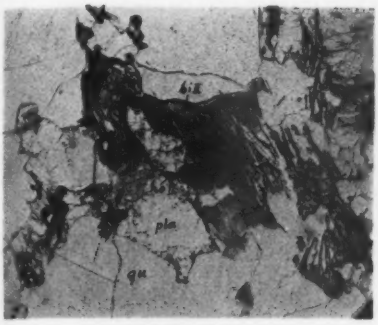


FIGURE 1

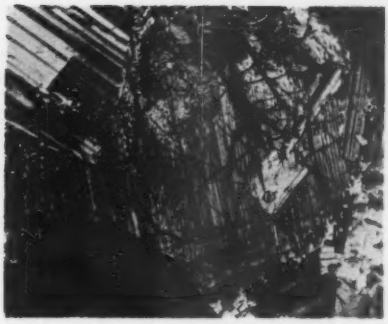


FIGURE 2

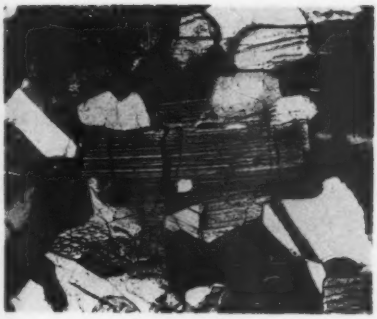


FIGURE 3

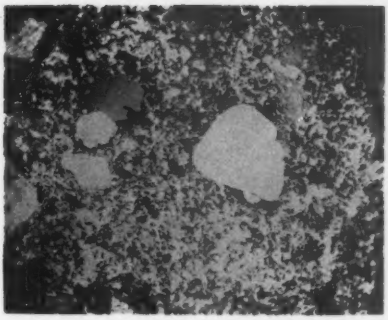


FIGURE 4

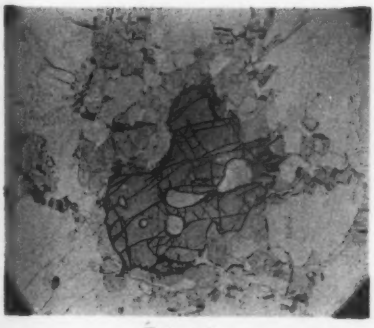


FIGURE 5

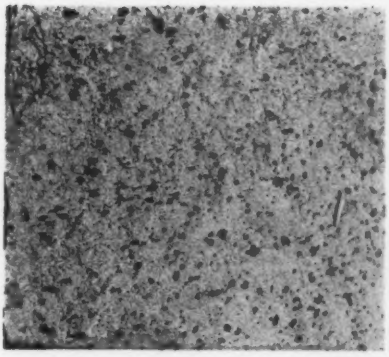


FIGURE 6

PHOTOMICROGRAPHS